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Paleogeography of the Coniacian–Maastrichtian Ages of the Late Cretaceous

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Abstract—Global lithologic–paleogeographic maps were compiled for the Coniacian, Santonian, Campanian, and Maastrichtian ages of the Late Cretaceous. Main features of paleogeography of continental and oceanic hemispheres of the Earth and ocean–continent transition zones during the second half of the late Cretaceous are considered. Continental hemisphere was divided into two asymmetrical segments: the northern segment of the Laurasian continent and the southern one, comprising isolated Gondwanan blocks. In the terminal Late Cretaceous, the sublatitudinal Tethys was divided, after the formation of integrated Atlantic Ocean, into two autonomous parts: the Caribbean basin in the west and the Late Cretaceous Tethys in the east. The oceanic hemisphere was occupied by the Pacific, whose central part represented a deep pelagic zone where abyssal basins widened and deepened. Most significant global paleogeographic transformations occurred in the peripheral zones of the Pacific and Tethys and also in continents bounding these basins.

Key words: *paleogeography, Coniacian, Santonian, Campanian, Maastrichtian.*

INTRODUCTION

The paper is devoted to a consideration of global lithologic–paleogeographic maps first compiled for the Coniacian, Santonian, Campanian, and Maastrichtian ages of the Late Cretaceous (Figs. 1–4). It continues the brief characterization of global maps compiled for every age of the Cretaceous to reveal sedimentation settings, paleogeography, and paleoclimatic zoning during the epoch of the warm biosphere. In our previous publications (Zharkov *et al.*, 1995, 1997), lithologic–paleogeographic maps compiled for the Berriasian, Valanginian, Hauterivian, Barremian, Aptian, Albian, Cenomanian, and Turonian ages of the Late Cretaceous were discussed. All the maps, including those presented in this paper, were compiled using the same approach, which was described earlier (Zharkov *et al.*, 1995).

The main principle of our methodical approach is to use the autonomous legends for each of five paleogeographic realms that existed during the Cretaceous: (1) pelagic zones of oceans; (2) continental slope/island arc foot and peripheral zones of oceans; (3) shelf and epicontinental seas; (4) humid zones and areas of continents; and (5) arid sea basins in interior and coastal areas of continents.

As a result, continental and oceanic regions, as well as transition zones between them, were distinguished. This also allowed us to discriminate land areas, shelf and epicontinental seas, and humid settings on continents, as well as central pelagic, peripheral hemipelagic, and island-arc zones in oceans. Magmatic and sedimentary–volcanogenic rock complexes are recognized according to their formation in either continental or oceanic settings. Magmatic rock associations of marginal–continental and intracontinental settings are distinguished on continents, whereas those of mid-oceanic ridges, intraplate seamountains, islands, plateaus, and island arcs are defined in oceans.

Compiled maps demonstrate peculiar features of the general distribution of basins with carbonate, terrigenous, glauconite-bearing, phosphate-bearing, black shale, evaporitic, chalk, and coal accumulation, as well as settings of bauxite and kaolinite formation on continents and in shelf and epicontinental seas during every age of the Cretaceous. They also provide an opportunity to clarify general patterns of distribution of hemipelagic, pelagic, and anoxic black shale deposits in oceans, and also of magmatic and sedimentary–volcanogenic complexes in oceans, continental margins and continents.

As noted in previous publications, it is impossible to consider paleogeographic and paleoclimatic peculiarities in the distribution of all the distinguished sedimentation and volcanism settings for separate ages, including those of the second half of the Late Cretaceous, in one paper. Therefore, when analyzing the paleogeography of the Cretaceous, one should pay attention only to some aspects of a principal significance. In this paper, we consider only two issues:

(1) main features of global paleogeography of the second half of the Late Cretaceous;

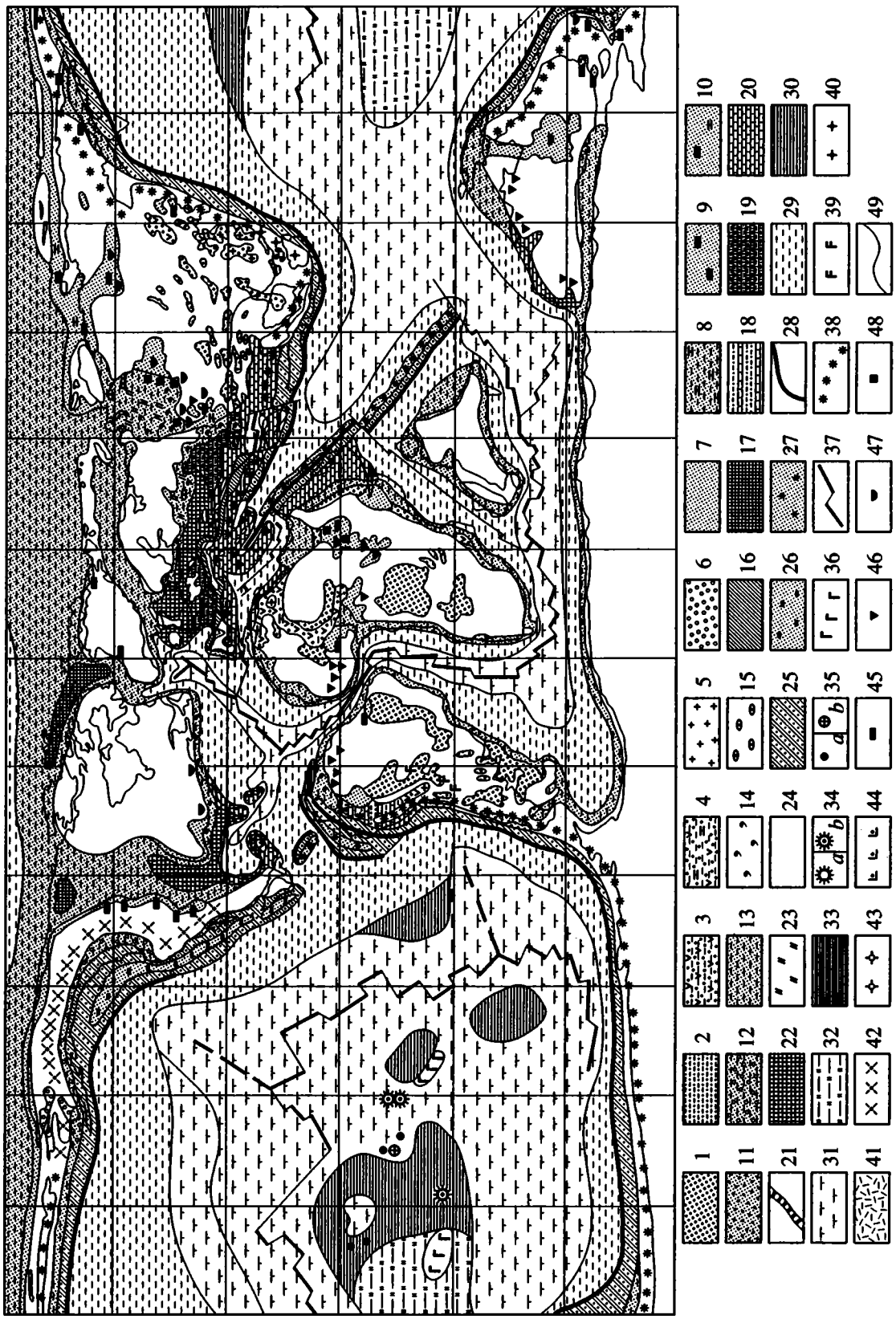
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Fig. 1. Lithologic–paleogeographic map for the Coniacian Age of the Late Cretaceous: (1–5) deposits of alluvial–proluvial plains, intramontane depressions, lakes, sebkhas, and lagoons in arid zones: (1) red-colored conglomerate, gravelstone, and sandstone; (2) red-colored and variegated sandstone, siltstone, and claystone; (3) sandstone, siltstone, and claystone with gypsum (anhydrite); (4) gypsum-bearing calcareous and terrigenous–calcareous; (5) salt-bearing; (6–11) deposits of alluvial and lacustrine–swampy plains, intramontane depressions, maritime plains intermittently flooded by sea, and lagoons in humid zones: (6) gray-colored conglomerate, gravelstone, and sandstone; (7) gray-colored sandstone, siltstone, and clay; (8) calcareous–terrigenous; (9) terrigenous coal-bearing (intracontinental); (10) calcareous–terrigenous coal-bearing (coastal); (11) terrigenous red-colored carbonate-free; (12–23) deposits of shelf and epicontinental seas: (12) conglomerate, sandstone, siltstone, and claystone; (13) sandstone, siltstone, and claystone; (14) glauconite-bearing; (15) phosphorite-bearing; (16) turbidites of shelf slopes and back-arc basins; (17) carbonaceous clayey, clayey–calcareous, calcareous–siliceous (black shales); (18) siltstone, claystone, and limestone; (19) clayey limestone, marl; (20) calcareous shallow-water (carbonate platforms); (21) reefs; (22) chalk; (23) opoka; (24) land; (25–29) deposits of the continental slope and rise, island arcs, and peripheral areas of oceans: (25) turbidites; (26) terrigenous, on rises in peripheral areas of oceans; (27) calc-alkalic and tholeiitic, as well as terrigenous–volcanogenic island-arc complexes; (28) sediments of deep-sea trenches; (29) hemipelagic clayey, calcareous–clayey, calcareous; (30–37) deposits of central areas of oceans: (30) pelagic clayey and calcareous–clayey (brown clays); (31) pelagic calcareous and siliceous–calcareous; (32) pelagic siliceous; (33) carbonaceous clayey, calcareous–clayey, siliceous–calcareous, and siliceous (black shales); (34) intraoceanic islands: (*a*) carbonate atolls and (*b*) volcanic alkalic and bimodal associations; (35) guyots: (*a*) exposed with phosphate and ferrous–manganese hardgrounds and (*b*) covered with pelagic calcareous ooze; (36) intraplate alkalic and tholeiitic basalts; (37) tholeiitic basalt of mid-ocean ridges; (38) calc-alkalic magmatic rocks of continental–marginal volcanoplutonic associations; (39) intracontinental alkalic and tholeiitic basalts and bimodal associations; (40) degrading volcanic belts; (41) distal ash; (42) degrading plutonic belts; (43) granitoids in orogenic belts; (44) gabbroids and basaltic rocks in orogenic belts; (45) coal-bearing deposits; (46) bauxite; (47) kaolinite clay; (48) iron ore; (49) boundaries.

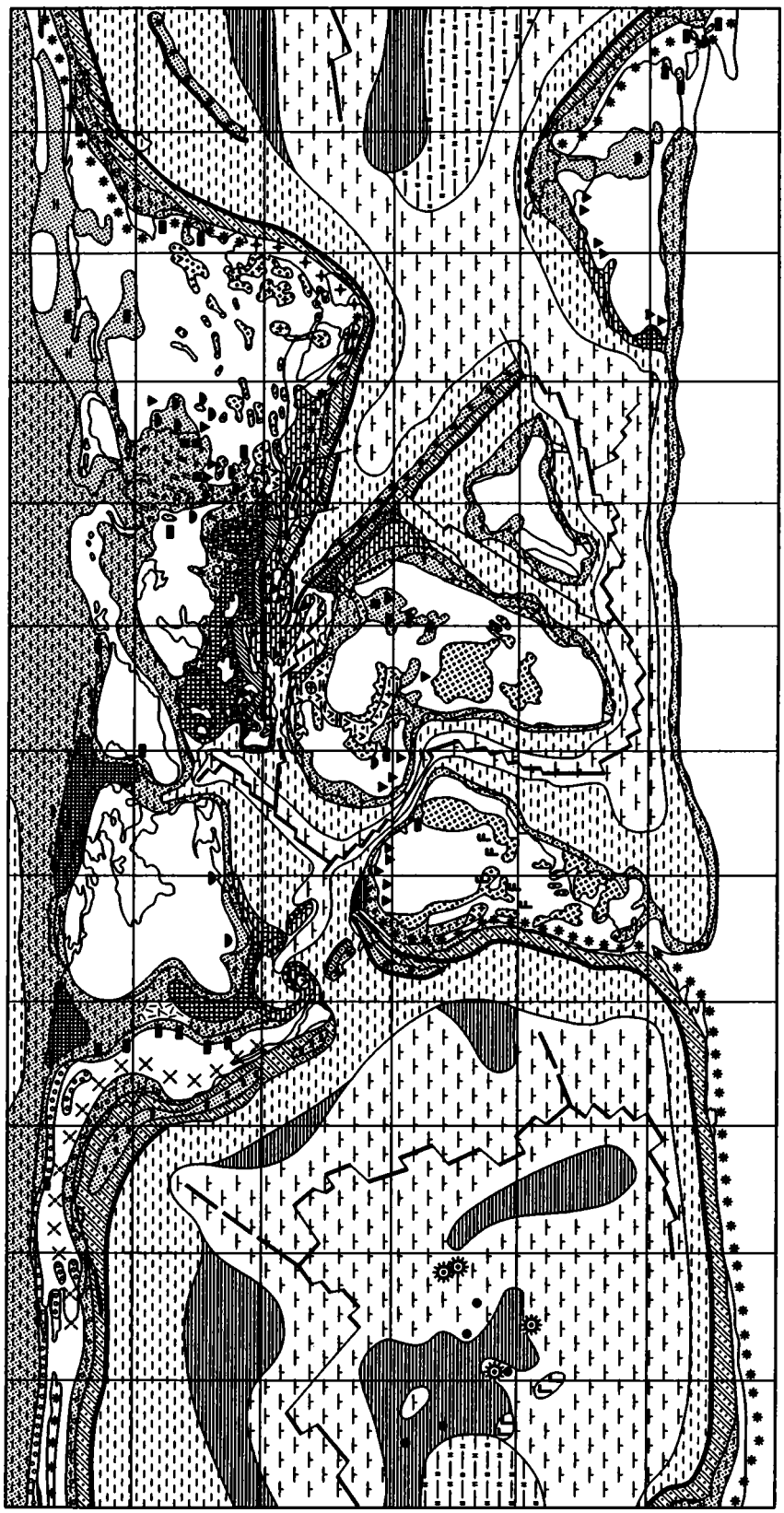

Fig. 2. The lithologic–paleogeographic map for the Santonian Age of the Late Cretaceous (symbols as in Fig. 1).

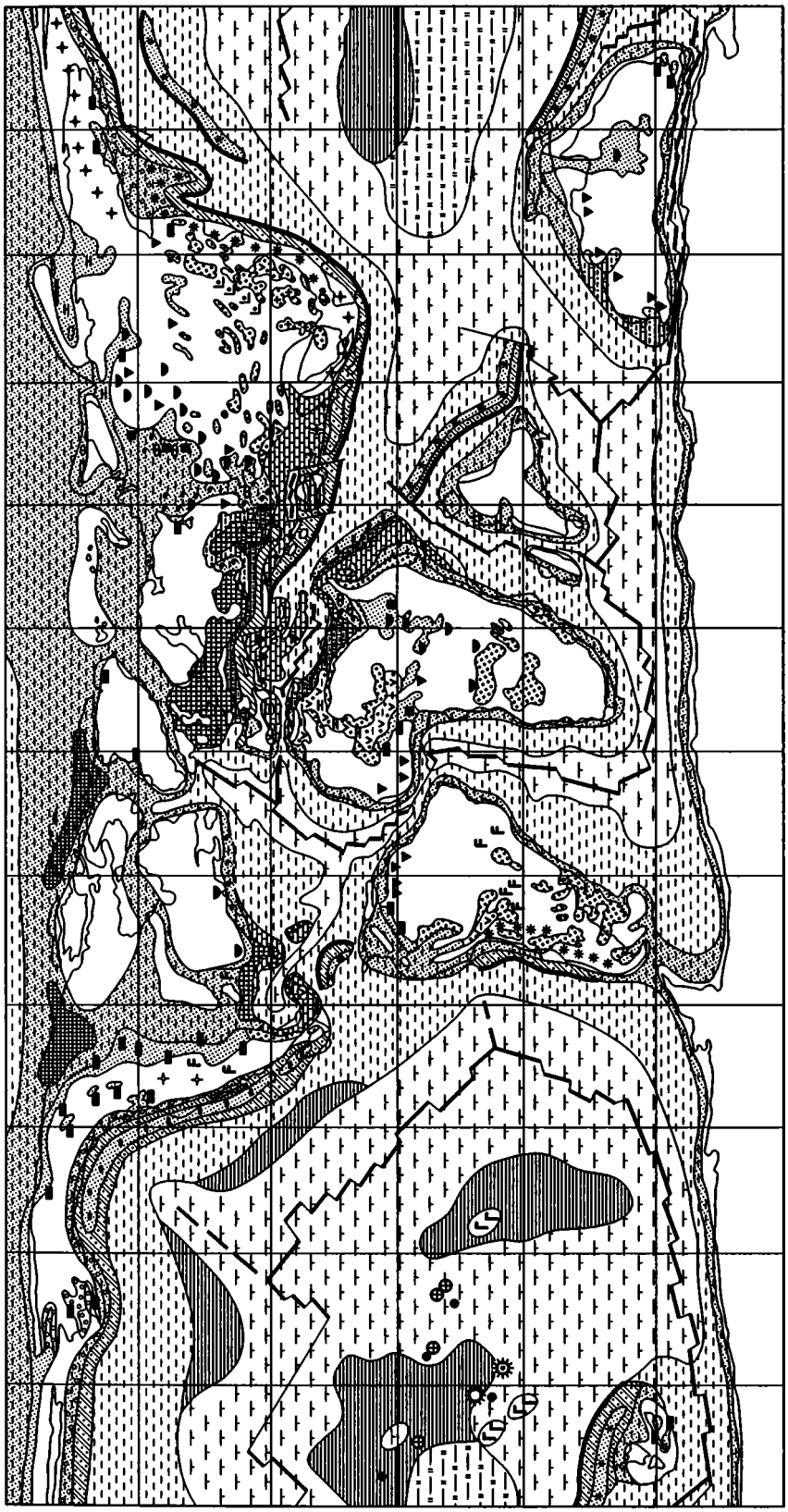

Fig. 3. The lithologic-paleogeographic map for the Campanian Age of the Late Cretaceous (symbols as in Fig. 1).

Fig. 4. The lithologic–paleogeographic map for the Maastrichtian Age of the Late Cretaceous (symbols as in Fig. 1).

(2) main regularities in distribution of paleogeographic settings in the continent/ocean transition zone.

Most of cited publications that were used when compiling lithologic–paleogeographic maps were listed in previous works (Zharkov *et al.*, 1995, 1997). The reference list of this paper includes only newer, additional works.

MAIN FEATURES OF PALEOGEOGRAPHY DURING THE CONIACIAN–MAASTRICHTIAN

The second half of the Late Cretaceous represents a transition period from the Mesozoic to Cenozoic planetary paleogeographic situation. It was a period when the Earth's surface was still divided into oceanic hemisphere occupied by the Pacific and continental hemisphere comprising all the continental blocks: Laurasia, South America, Africa, Hindustan, and Eastern Gondwana, which included Antarctica, Australia, and New Zealand at the beginning of the second half of the Late Cretaceous. The global paleotectonic and paleogeographic asymmetry of the Earth was responsible for peculiarities in many paleogeographic reorganizations that occurred in oceanic and continental hemispheres throughout this period.

The continental hemisphere was characterized by a mosaic distribution of continents separated by oceans of sublatitudinal and submeridional strikes. This hemisphere retained a paleogeographic asymmetry typical of the entire Cretaceous. The asymmetry was reflected in the opposite position of the huge Laurasian continent, located mainly in temperate and high latitudes of the northern hemisphere, on the one hand, and in the isolated Gondwanan continental blocks of South America, Africa, Hindustan, and Eastern Gondwana, situated mostly in the southern hemisphere, on the other. The Tethys ocean stretching in the sublatitudinal direction along the northern tropical and subtropical belts was a structure separating these asymmetric paleogeographic segments. The processes that controlled initiation and opening of new oceans and interrelated breakup events and movements of continental blocks continued in the southern hemisphere.

In the southern hemisphere, the progressing extension of the South Atlantic and South Ocean occurrred throughout the entire second half of the Late Cretaceous (Sclater *et al.*, 1977; Rabinowitz and La Brecque, 1979; Barron *et al.*, 1981; Ziegler *et al.*, 1982; Emery and Uchupi, 1984; Zonenshain *et al.*, 1984, 1990; Krasheninnikov and Basov, 1985; Barron, 1987; Scotese *et al.*, 1988; Sager and Scotese, 1989; Scotese, 1991; Zonenshain and Kaz'min, 1992; Khain and Balukhovskii, 1993; Golonka *et al.*, 1995). The width of the South Atlantic increased by approximately 2000 km (from 3500 km in the Coniacian to more than 5500 km in the Maastrichtian) in extreme southern areas and by 1200 km in its northern areas, where it became 2500 km wide in the Maastrichtian. Simultaneous extension of the North Atlantic improved connections between South and North Atlantic and resulted in formation of the intergrated Atlantic Ocean of the submeridional strike. The South Ocean stretched near 60°S from the South America on the west to Australia on the east occupying areas between Antarctica, Africa, and Hindustan. Its widths varied from 2000–2500 km to 3000–3500 km. In its central part, the South Ocean was connected with the Eastern Tethys via the deep Mozambique–Somalia seaway, and, in the eastern part, it freely opened into the Himalayan–Indonesian Tethys and Pacific.

In the Campanian, Australia was separated from Antarctica due to southeastern and eastern progradation of the spreading axis, which resulted in the formation of another deep seaway between the Pacific and South oceans (Veevers, 1984; Patriat and Segoufin, 1988; Scotese *et al.*, 1987, 1988; Scotese, 1991). The separation of New Zealand from Australia and Madagascar from Hindustan also occurred at that time.

While the continental breakup with initiation and opening of deep oceanic basins proceeded in the southern segment of the continental hemisphere, radial subsidence continued in the northern segment of Laurasia, where simultaneous formation of epicontinental seas of submeridional or similar strike divided the Laurasia continent into autonomous land blocks. In the west, the Western Interior sea basin connecting the Caribbean part of the Tethys with the Amerasian ocean continued to exist throughout the entire time period in question; it separated the land area of the Rocky Mountains, stretching along the western periphery of the North American margin of Laurasia, from the Canadian lowland. In interior areas of Laurasia between the Canadian lowland and Greenland and also between Greenland and Baltia, the Labrador–Baffin and Norwegian seaways connected the North Atlantic with the Amerasian ocean. In the east, the Laurasia continent always hosted the inland West Siberian sea, intermittently connected, via the Turgai seaway, with the Tethys-related epicontinental seas. Eastern Laurasia was occupied by a spacious land, Angarida, whose eastern and southern margins hosted mountain systems, whereas northern and western areas were occupied by lowlands. From the Coniacian to Maastrichtian ages, huge northern and southern areas of Laurasia were occupied, along with above-mentioned seaways, by spacious epicontinental seas. Its southern margins were covered by the interconnected North, Polish–Lithuanian, Central European, East European, Turan, and Afghan–Pakistan seas, whereas the Sverdrup, North Greenland, Barents, Kara, and other shelf and epicontinental seas existed along the northern margin. More than 40% of Laurasia was covered by seas. In the paleogeographic respect, this feature substantially distinguished this continent from all southern continental blocks of the Gondwana group, a significant part of which constantly represented land areas. Only the Trans-Sahara seaway similar, in some features, to Laurasian seaways existed in North Africa.

However, the Trans-Sahara seaway was located in the tropical and equatorial zones, and this resulted in a combination of evaporitic and normal marine sedimentation. During the second half of the Late Cretaceous, paleogeographic patterns of the continental hemisphere resembled the situation that appeared beginning in the mid-Cretaceous and was related to the asymmetric position of epicontinental and shelf seas in Laurasia, on the one hand, and in southern continents, on the other.

According to recent paleogeographic reconstructions (Philip *et al.*, 1993a, 1993b; Camoin *et al.*, 1993a, 1993b), the Late Cretaceous Tethys was subdivided into five main region replacing one another in the west–east direction: (1) the Caribbean, located between two open oceans, the Pacific on the west and the Atlantic Tethys on the east; (2) the Atlantic, which represented a symmetric oceanic basin with the mid-ocean ridge; (3) the Mediterranean, consisting of the mosaic system of shallow carbonate platforms separated by deep sea basins often floored by oceanic crust; (4) the Eastern Tethys, comprising oceanic areas between Eurasia and the Arabian margin of Africa; and (5) the Himalayan–Indonesian Tethys, which covered areas between the Asian part of Laurasia on the north and Hindustan and Australia on the south. At the same time, one should remember that a series of important paleogeographic events, which resulted in principal changes in configuration and extension of the Cretaceous Tethys, occurred during the second half of the Late Cretaceous, particularly in the Campanian. They were related to the transformation of Atlantic basins into the integrated ocean extending in submeridional direction. In every paleotectonic and paleogeographic respect, the North Atlantic was united with the South Atlantic into a single planetary-scale ocean at that time. The North Atlantic up to 5000 km wide represented an oceanographic barrier that isolated the Caribbean realm from the Mediterranean province (Klitgord and Schouten, 1986; Camoin *et al.*, 1993b). That was the reason why the North Atlantic was no more a constituent of the Tethys, but became an element of the Atlantic Ocean, which divided the sublatitudinal Cretaceous Tethys into two isolated and autonomous parts unequal in size: the western Caribbean part and the eastern one comprising the Mediterranean, Eastern, and Himalayan–Indonesian Tethys. Beginning from the terminal Cretaceous, only this eastern part should be considered as the Tethys proper.

Despite the Tethys fragmentation, the more-or-less free water exchange with the Pacific continued to exist in the east and west, and gave rise to circum-global Western currents in tropical latitudes of the northern hemisphere (Luyendyck *et al.*, 1972; Gordon, 1973; Berggren and Hollister, 1974; Camoin, 1993). It is assumed that circular clockwise surficial currents were quite significant in the North Atlantic, Mediterranean, and Eastern Tethys (Barron and Peterson, 1989; Tucholke and McCoy, 1986). It should be noted that the free influx of warm tradewind currents from the Pacific into the South Ocean and further into the South Atlantic occurred via the wide seaway between Hindustan and Australia and through the Mozambique–Somalia strait. As for the surficial water exchange with the Amerasian ocean, it is assumed that respective currents passed through the Labrador–Baffin and, probably, Norwegian seaways (Berggren and Olson, 1986; Gradstein *et al.*, 1991).

The Pacific hemisphere of the Earth represented a deep oceanic basin, which had free exchange with the Tethys and South Ocean via wide seaways in the west and restricted connection with the Caribbean province and Atlantic Ocean in the east. In the tropical latitudes, the circular western current passed through the entire Pacific. It is presumed that spacious anticyclonic circulation existed to the south and north of the tropical currents.

The major central part of the Pacific Ocean bottom was occupied by the Pacific plate, which gradually extended owing to the spreading. The Kula (on the north), Farallon (on the east), and Phoenix (on the southeast) plates bounding the Pacific were reduced in area as a result of subduction. According to paleotectonic reconstructions (Scotese *et al.*, 1988; Kononov, 1989; Pushcharovskii and Melankholina, 1992) that were modified by M.V. Kononov in our maps, the sublatitudinal part of the spreading ridge between the Pacific and Kula plates migrated northward from 10°–20°N in the Coniacian to 25°–35°N in the Maastrichtian. The submeridional part of the ridge migrated eastward nearing the position of the modern East Pacific Rise at the end of the Cretaceous. Correspondingly, intraplate morphostructures (undersea mountains and rises, deep-sea basins) located on the Pacific plate also migrated together with their facial complexes in the course of spreading. Simultaneously, lithospheric plates with related bottom morphostructures experienced the isostatic subsidence.

During the entire time period in question, the central part of the Pacific (including the entire Pacific plate) reprsented a spacious pelagic area, which was characterized by extremely low terrigenous influx from continents, low bioproductivity of surficial waters, and active aeration of bottom waters. Under such conditions, red (oxidized) pelagic sediments were slowly accumulated: calcareous nannoplanktonic–foraminiferal oozes in the undersea rises above the carbonate compensation depth (CCD) and siliceous (radiolarian) to clayey deposits (pelagic brown clays) in abyssal plains below the CCD (Murdmaa, 1987). The CCD level, i.e., the boundary between the carbonate and carbonate-free facis, which was highest in the mid-Cretaceous, slightly lowered by the end of this period. Correspondingly, carbonate sediments prograded to greater depths.

From the Coniacian to Maastrichtian, abyssal areas (below the CCD, i.e., 4–5 km deep) of the Pacific and, probably, neighboring oceanic plates gradually deep-

ened and widened that extended areas of carbonate-free clayey and siliceous–clayey sedimentation. In addition to Central Basin, the South Basin was formed, and brown pelagic clays, the indicators of low bioproductivity zones ("oceanic deserts"), were deposited here. In the Central Basin, which gradually moved northward, two facial complexes, pelagic clayey and clayey–radiolarian, were accumulated. Some data indicate that siliceous sediments were mostly confined to the southwestern equatorial part of the basin, where a sublatitudinal belt of silica accumulation, probably related to the above-mentioned circum-tropical currents, was formed before the Maastrichtian.

In the intraplate rises and mid-oceanic ridges (above the CCD), pelagic sediments accumulated.

The northward migration of the Mid-Pacific Mountains was in progress, and they reached the equatorial zone and subsided below sea level to depths of several hundreds meters to be transformed into guyots. Bottom currents washed the guyot tops and created environments of nondeposition or extremely slow pelagic sedimentation (strongly condensed sequences) and phosphate and ferromanganese hardgrounds (Murdmaa *et al.*, 1995). The volcanic activity continued south of the Mid-Pacific Mountains, in the Darwin swell, and resulted in the formation of new seamounts of the modern archipelago of the Marshall Islands.

MAIN REGULARITIES IN DISTRIBUTON OF PALEOGEOGRAPHIC SETTINGS IN THE CONTINENT–OCEAN TRANSITION ZONES

At the end of the Cretaceous (88–65 Ma), paleogeographic settings (and corresponding lithologic complexes) of the continent–ocean transition zone were primarily controlled by the interaction between oceanic and continental plates. For instance, relatively slow widening of the Atlantic (Zonenshain and Kuz'min, 1992) favored the quiet and stable tectonic regime in its continental framing and the formation of a relatively narrow band of shallow-water terrigenous deposits along the periphery, which graded seaward into hemipelagic clayey and carbonate–clayey facies. By contrast, in the Pacific and Tethys, converging oceanic and continental plates produced quite different landscapes of the continent–ocean transition zone.

Continental marginal zones of the Pacific had a complicated structure. In the second half of the Late Cretaceous, the northeastern peripheral part of the Pacific was characterized by a combination of strike-slip movements and subduction of oceanic plates under the North American continent (Csejtey *et al.*, 1982; Monger *et al.*, 1982; Coney, 1987; Frazier and Schwimmer, 1987; Plafker *et al.*, 1989; Wallace *et al.*, 1989; Haugerud *et al.*, 1991; Livaccari, 1991; Underschultz and Erdmer, 1991; Ingram and Hutton, 1994). A series of narrow undersea mountain chains with relatively shallow-water terrigenous sedimentation was formed along the peripheral zone of the ocean during the Coniacian–Santonian (Figs. 1, 2). They represented terranes, including the most spacious Talkeetna terrane. The mountain chains were bounded by troughs with turbidite accumulation in both continental and oceanic sides. Eastern structures of the last type are usually considered as marginal troughs relative to the North American continent. In the Wrangellia and Talkeetna mountains, the thickness of the Coniacian–Campanian flysch is up to 1000 m (Plafker *et al.*, 1989). Similar flysch deposits are also known from southern areas, e.g., from western California (Khain and Balukhovskii, 1993). Subsequent strike-slip movements resulted in the migration of the Talkeetna terrane and corresponding mountain chain (Figs. 1–3). In the Campanian, the latter joined the western margin of North America, which resulted in the formation of thick (800–2000 m) silty–sandy coal-bearing deposits intercalated with conglomerate beds near the continent edge.

The western margin of the North American continent was characterized in the second half of the Late Cretaceous by a permanent uprising and formation of the Cordillera orogenic system. Two main factors were responsible for this last event: the rise of the continental margin and intense eastward horizontal movements of rock masses (the Laramian orogeny). These movements resulted, first, in the thrust–nappe structure of mountains and, second, in asymmetric patterns of distribution of continental and coastal basins along the eastern Cordillera piedmont. Orogeny was most intense during the Campanian–Maastrichtian, as it is reflected in deposits of synorogenic basins adjoining the Cordilleras in the east (Coney, 1987). The area of terrestrial sedimentation between the Cordilleras and Western Interior seaway was of a complex structure (Dickinson *et al.*, 1988). Three zones are distinguished in the Rocky Mountains located within this area. The western zone that directly joined the Cordilleras included the so-called barrier basins, where sedimentation was most intense and long-lasting. The intermediate zone consisted of alternating ridges and depressions. The third eastern zone included peripheral basins. In the Coniacian–Maastrichtian, small but numerous basins of terrestrial sedimentation existed along strike-slip faults in the Alaskan Cordilleras (Fisher *et al.*, 1982). Thick Upper Cretaceous terrigenous sequences filling these basins incorporated coal-bearing members, whose abundance was particularly significant at the Cretaceous–Paleogene boundary (Fig. 4).

Inasmuch as the intensity of subduction sharply decreased in the interval from the Turonian to Campanian and strike-slip movements became prevalent, the type of magmatism also changed in the west of North America. As shown earlier (Zharkov *et al.*, 1995; Filatova, 1996), in the Albian–Cenomanian, this area included the lateral structural succession typical of the active continental margin: a deep-sea trench marking the subduction zone; a fore-arc trough with turbidite

accumulation; and a plutonic belt of the continental margin. The second half of the Late Cretaceous was marked by the gradual cessation of activity in this belt (Figs. 1, 2). Judging from discrete patterns of magmatic rock ages (Evenchick, 1991), the magmatic activity was pulsating against the background of constant continental margin rising and displacement of rock sequences eastward along overthrust surfaces and parallel to the continent–ocean boundary along strike-slip faults.

The formation of the Laramide orogenic belt and corresponding Cordillera system in the Campanian–Maastrichtian changed the type of magmatism and resulted in the appearance of muscovite granitoids in California (Karlstrom *et al.*, 1993). Further development of the Laramian orogeny was accompanied by basaltic volcanism in the strike-slip zones (Dickinson *et al.*, 1988; Plafker *et al.*, 1989) (Fig. 4).

Three segments (northern, central, and southern), which show certain differences in their evolution, can be distinguished in the southeastern peripheral area of the Pacific and in the west of South America during the Late Cretaceous (Herve *et al.*, 1987; and others).

The most peculiar history was that of the Venezuela–Ecuador segment, where the synchronous formation of perioceanic island-arc and continental–marginal volcanic belts occurred in the Late Cretaceous (McCourt *et al.*, 1984; Megard, 1987; Aguirre *et al.*, 1992; Van Thrournout *et al.*, 1993; Desmet, 1994; and others). During the Laramian orogeny, which commenced in the Campanian and continued till the early Paleogene (Megard, 1987), the island-arc belt was obducted onto the continent. This resulted in crustal thickening of the northwestern margin of the South American continent and the formation of the Laramian thrust–nappe orogenic belt comprising the huge structure of the Andes. The eastern front of the continental margin was occupied by the synorogenic basin with terrigenous sedimentation. During the Campanian–Maastrichtian, the northwestern margin of the South American continent was involved in the marine sedimentation cycle lacking any significant volcanism (Figs. 3, 4).

The central segment of western South America hosted the Araukan continental–marginal volcanic belt (Lomize, 1983), whose formation commenced after the Peruvian (Austrian) orogeny. Continental volcanic deposits in the central segment of this belt are up to 4000–6000 m thick, narrowing northward down to 1000 m and less. During the Early–Late Mesozoic period, the axis of continental–marginal volcanic belt migrated eastward (James, 1971; Coira *et al.*, 1982; Hevre *et al.*, 1987) in response, as usually explained, to the gradual flattening of the subduction zone plane. Westward of the central segment of the Andes, the Coastal Range zone of uplifts and the fore-arc trough with terrigenous–volcanogenic deposits were located. In the east, the volcanic belt was in contact of a zone of uplifts, which, in turn, joined, further to the east, the submeridional chain of back-arc basins with terrestrial deposits and volcanics. The latter structures were represented by the Subandean and Andean basins of Bolivia and Peru, as well as by the Chacoparana and Salta basins of Argentine (Grier *et al.*, 1991).

The continental–marginal volcanic belt of the central segment evolved through the entire Late Cretaceous. In the Coniacian–Santonian, it extended into the southern segment of South America and further to the south along the Antarctica margin. The southern segment, to the east of the volcanic belt, hosted the mountain chain that was the eastern boundary of the Magellan marginal basin. This zoning in the southern segment of western South America, which appeared in the late Albian, existed till the Campanian. The Laramian orogeny resulted in the general rise of the continental margin, in the cessation of volcanic activity and in the eastward migration of the Magellan basin (Wilson, 1991).

In general, the Late Cretaceous was marked by the development of two types of basins in the western margin of South America behind the volcanic belt: (1) the basins genetically related to mountainous chains migrating eastward and generally marked by terrigenous sedimentation; and (2) back-arc (relative to the volcanic belt) basins of rift origin with alkaline and bimodal magmatism. Basins of the first type were widespread along the entire western margin of South America, whereas local rift basins were mainly characteristic of the central segment of the South American continent during the Late Cretaceous.

The northwestern and western peripheral areas of the Pacific along the Asian continental margin hosted a chain of active volcanoes of an extended continental–marginal volcanic belt with several diachronous segments: Okhotsk-Chukotka, Sikhote-Alin, and Cathasian. The last one (*Atlas* ..., 1985) included volcanic structures of the maritime part of China, together with those of southern Korea and northern Japan, now separated by the Sea of Japan (Filatova, 1990; Chang, 1995).

The continental–marginal volcanic belt was associated with a deep-sea trench (Figs. 1–3). The inner, continental slope of the trench was of a zoned structure. The zone near the continent was occupied by a shallow sea with abundant islands and island chains. During the early epochs of the Late Cretaceous, the sea accumulated flyschoid sequences, which later were replaced by coarse olistostrome–molasse complexes (Zinkevich, 1981; Filatova, 1995, 1996). Futher seaward, thick turbidite sequences were formed on the deeper slope of the trench. They are exemplified by the Upper Cretaceous sequences of western Sakhalin (Rozhdestvenskii, 1987; Zonenshain *et al.*, 1990) and by deposits of the Simanto and Hidaka zones of Japan (Taira, 1985; Kimura *et al.*, 1990). An ensemble of volcanic island arcs and marginal seas was located in the perioceanic zone (Bogdanov and Filatova, 1988; Zonenshain *et al.*, 1990; Sokolov, 1992); one of its segments was repre-

sented by the Irunei island arc. This structural ensemble constantly migrated landward and joined the continent in the Maastrichtian (Mitrofanov, 1977; Alekseev, 1979; Bogdanov *et al.*, 1982; Filatova, 1988; Bogdanov and Til'man, 1992; Parfenov *et al.*, 1993). Fragments of its island arcs are preserved in the form of tectonic nappes in Kamchatka and eastern Sakhalin (Pushcharovskii *et al.*, 1983; Pushcharovskii and Melankholina, 1992; and others), as well as in Hokkaido (Taira, 1985, Kimura and Kensaku, 1986; Maruyama and Seno, 1986; and others).

The considered events occurred during the Laramian orogeny, which was characterized by the general rise of the Asian margin (Zinkevich, 1981; Ruzhentsev *et al.*, 1982; and others), where isolated basins accumulated coal-bearing molasses. This orogeny caused cessation of volcanism in some segments of the continental–marginal volcanic belt (for instance, in the Okhotsk-Chukchi belt), although other segments continued to be active till the beginning of the Paleogene.

Reorganization of oceanic plates in the northwestern and western parts of the Pacific at 85–74 Ma was noted by many researchers (Kimura *et al.*, 1985, Maruyama and Seno, 1986; Jolivet *et al.*, 1988; Zonenshain *et al.*, 1990; Khain and Balukhovskii, 1993; Parfenov *et al.*, 1993; Okada and Sakai, 1993; and others). In the Campanian–Maastrichtian (Figs. 3, 4), the Sea of Okhotsk represented an uplifted area occupied by a shallow sea with island chains and bounded by the volcanic belt in the south. (Maruyama and Seno, 1986). In the continental margin, this time period was marked by the formation of pull-apart basins, which accumulated terrigenous coal-bearing deposits. Such basins, which are often of a half-graben form, were long known in China and recently were also discovered in Japan (Okada and Sakai, 1993). Movements along faults that confine these basins resulted in the formation of relatively small, but abundant basaltic plateaus (Filatova, 1988).

As for the southwestern periphery of the Pacific, important events occurred there in the terminal Late Cretaceous. Whereas the Coniacian–Santonian time was characterized by the formation of the volcanic belt on the Australian–Antarctic margin (Howell, 1980), the Campanian was marked by the opening of the oceanic basin between Australia and Antarctica and by the formation of the Tasman Sea bounded by the volcanic island arc on the Pacific side (Veevers, 1984; Scotese *et al.*, 1988) (Figs. 3, 4).

Drastic changes in the plate movement occurred in the Tethyan region during the Late Cretaceous (Dercourt *et al.*, 1985, 1993; Zonenshain *et al.*, 1987). The Eurasian and African–Arabian blocks have exchanged their previous southward migration for northward movements. The Bay of Biscay opened, and the Apulian block was separated from Africa. The oceanic plate of the Eastern Carpathian and Balkan basins subducted southward to give rise to the formation of volcanic belts. The Late Cretaceous subduction of the oceanic crust under Eurasia resulted in formation of the major South Eurasia volcanic belt along the northern margin of the Tethys (Kaz'min *et al.*, 1987). The western extremity of the latter is recognized in the Bulgarian Highland. Eastward, the belt is traced in the Western Pontids and further to the south, in the echelon-like structures of the Eskisehir Mountains, where volcanism commenced in the Turonian (Sengor and Yilmaz, 1981). In the Transcaucasian segment of the volcanic belt, a thick volcanic sequence was formed in the latest Albian–early Campanian, whereas Campanian–Maastrichtian volcanics are much less widespread here. In general, two stages of the volcanic activity can be distinguished in the Pontian–Transcaucasian segments of the belt: the Albian–early Campanian and late Campanian–Maastrichtian. The main volume of volcanics was accumulated during the first stage. The volcanics are referred to calc-alkaline–alkalic and, less commonly, to tholeiitic magmatic series (Lordkipanidze, 1980). The second stage was marked by basaltic eruptions of subalkalic composition (Kaz'min *et al.*, 1987); this allows an assumption that the last stage was not subduction-related and induced after cessation of subduction during the Laramian orogeny.

The eastern continuation of the Late Cretaceous belt is traced in the northwestern Elburs region. Further to the east, calc-alkaline volcanics are recorded in tectonic fragments of the Mekran ophiolites between the Dgaz-Murian basin in the north and Badgan-Dur-Kan zone in the south. Still further to the east, volcanics of the Late Cretaceous age occur in Afghanistan.

At the beginning of the Coniacian Age, the Cyprus ensimatic island arc appeared in the southern part of the Tethys (Dercourt *et al.*, 1985; Knipper, 1985; Zonenshain *et al.*, 1987). In the Indian Ocean, the Kohistan (Bard, 1983), or Dras (Dietrich *et al.*, 1983) intraoceanic island arc with associated deep-sea trench in the south was formed in the time period of 90–75 Ma (Figs. 1, 2). Tholeiites erupted in the arc (Dietrich *et al.*, 1983) and volcanogenic–terrigenous deposits were accumulated in the fore-arc basin (Robertson and Degnan, 1994).

In the terminal Cretaceous time and at the Cretaceous–Paleogene boundary, the Tethyan region experienced intense orogeny (synchronous with the Laramian one), which resulted in the closing of the southern part of the Eastern Tethys and in a collision between Africa and Eurasia (Dercourt, 1985, 1993; Zonenshain *et al.*, 1987; Knipper and Sharas'kin, 1995). Sedimentary sequences of the Alpine and Carpathian–Pannonian basins were deformed. The Late Cretaceous volcanism in the Carpathians ceased. At the end of the Late Cretaceous (approximately 80 Ma), the African–Arabian margin was subducted under the Cyprus arc and ophiolite nappes were obducted onto the continental margin. The Campanian was marked by a collision of the Kohistan island arc with the northern margin of India

(Bard, 1983; Zonenshain *et al.*, 1984). In the modern structure, tectonic nappes of island-arc volcanics and fore-arc basinal complexes occur in the Ind-Zangpo collision suture, which was formed in the Eocene (Allegre *et al.*, 1984).

The Late Cretaceous collision was accompanied by the gradual degeneration of the South Asian volcanic belt, whose activity had slowed down since the Campanian and almost completely stopped in the Maastrichtian (Fig. 3). Tectonic compression and obduction along the southern boundary of the Tethys were accompanied by substantial extension of back-arc basins in the Eurasian continental margin. These basins formed a system of Paratethyan seas 900 km wide and stretching from the Balkans to eastern Iran. Two (Western and Eastern) Black Sea basins opened, and the Central Iran oceanic basin between Lut, Sanandadg-Sirdgan, and Elburs structures increased in dimensions at that time.

CONCLUSION

The following main peculiarities of paleogeographic reorganizations that occurred during the second half of the Late Cretaceous can be inferred from the analysis of lithologic–paleogeographic maps compiled for the Coniacian, Santonian, Campanian, and Maastrichtian ages.

(1) Many features of paleogeographic reorganizations were related to the global paleotectonic and paleogeographic asymmetry of the Earth. The asymmetry was primarily reflected in the contraposition of oceanic and continental hemispheres, as well as in the contraposition of Laurasian continent in the north and isolated Gondwanan continental blocks of the south in the last hemisphere. The southern segment of the continental hemisphere was characterized by the development of oceans, which appeared in earlier epochs, and by origin of new oceanic basins (South Atlantic, South Ocean, Mozambique–Somalia and others) associated with break-up and migration of continental blocks (South America, Africa, Hindustan, Antarctica, New Zealand). Simultaneously, radial subsidence and formation of sublatitudinal epicontinental seas occurred in the northern segment of Laurasia. These asymmetric segments of the continental hemisphere were divided by the sublatitudinal Tethys Ocean. After integration of Atlantic basins into a single global structure at the end of the Late Cretaceous, the Tethys was diveded into two independent parts: the Caribbean basin in the west and Late Cretaceous Tethys proper in the east, where it included the Mediterranean, Eastern, and Himalayan–Indonesian parts. The oceanic hemisphere was occupied by the Pacific. The Pacific plate of its central part was gradually extending due to spreading, whereas the Kula, Farallon, and Phoenix plates were shrinking owing to subduction. During the entire second half of the Late Cretaceous, the central part of the Pacific represented a deep pelagic province with gradually extending and deepening abyssal plains.

(2) The paleogeographic situation in the ocean–continent transition zones was controlled by the interaction between oceanic and continental plates. Most significant paleogeographic reorganizations occurred in the peripheral zones of the Pacific and Tethys and also in continents around them.

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On the Upper Proterozoic Stratigraphy in the Kola Peninsula Coast (the Barents Sea Region)

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Abstract—The results of joint fieldwork with geologists from the Geological Survey of Norway refine the Upper Proterozoic correlation scheme for the Barents Sea region and substantiate recognition of three structural-facial zones: Terskii (a continental area), Tana–Kil'din (a stable continental margin), and Barents Sea–Rybachii (an active transitional area to a continental slope). There were three extensive stages of the Late Proterozoic sedimentation: (1) the Late Riphean accumulation of undifferentiated deposits at the edge of an active continental slope (the Kongsfjord Formation, the Eina and Bargoutnyi groups); (2) deposition of transgressive differentiated sediments, including carbonates, in the terminal Late Riphean (the uppermost beds of the Barents Sea Group, the Vadso and Tana groups, Kil'din Group, and a part of the Volokovaya Group); and (3) the Late Riphean–Vendian sedimentation marked by accumulation of tillite horizons in the Norwegian territory (the Lokvikfjellet and Vestertana groups and, probably, a part of the Volokovaya Group). The large lineament separating the Tana–Kil'din and Barents Sea–Rybachii zones is named "The Sokolov line" in honor of B.S. Sokolov, who was first to recognize this structural element as the northern border of the Russian platform in the Late Precambrian time.

Key words: *Barents Sea region, Kola Peninsula region, Upper Proterozoic, Riphean, Vendian, stratigraphy, correlation.*

The northwestern Murmansk area of Russia (the Srednii and Rybachii peninsulas, Kil'din Island, and the Terskii coast of the Kola Peninsula) is among geological objects of a great importance for the Riphean and Vendian stratigraphy of the North European coast. According to results of our long-term studies, we previously suggested Kil'din Group of the Srednii Peninsula and Kil'din Island to be older than the sedimentary succession of the Rybachii Peninsula (Lyubtsov *et al.*, 1989). This suggestion was consistent with the opinions of D.F. Agap'ev, T.P. Vronko, E.M. Lyutkevich, L.Ya. Kharitonov, and V.Z. Negrutsa (1971). In 1989–1993, we investigated the adjacent areas of Norway and Russia. The work was organized by the Geological Survey of Norway and the Geological Institute of the Kol'skii Research Center of Russian Academy of Sciences on the basis of a joint project, "The Northern Region." As a result, we revised our previous views and proposed a new scheme of subdivision and correlation of Upper Proterozoic deposits in the Kola Peninsula coast. The presented scheme is based on the revived idea of A.A. Polkanov, who assumed an older age of the Rybachii Peninsula sedimentary succession and its analogues with respect to Kil'din and Volokovaya groups, and also on the distinguished structural zoning of the Late Proterozoic complex in the Finnmark–Timan sector of the North European coast.

In the course of long-term (1970–1995) investigations of the Upper Proterozoic deposits of this region, a wide range of techniques was used, such as detailed field observations, lithologic–petrographic studies (Lyubtsov, 1975; Lyubtsov *et al.*, 1989), interpretation of aerial photographs (Lyubtsov *et al.*, 1978), geochemical analysis (Predovskii *et al.*, 1987; Lyubtsov *et al.*, 1981), and studies of microfossils (Lyubtsov, 1980; Lyubtsov *et al.*, 1989) and stromatolites (Krylov and Lyubtsov, 1976; Raaben and Lyubtsov, 1993; Raaben *et al.*, 1995). We hope this makes the suggested conclusions sufficiently substantiated.

DISTRIBUTION AND STRUCTURE OF UPPER PROTEROZOIC DEPOSITS IN THE KOLA PENINSULA COAST

Distribution area of Upper Proterozoic deposits in the Kola Peninsula coast corresponds to a minor part of the Riphean–Vendian zone of sedimentation in the region of the White and Barents seas. In the peninsula, weakly deformed Upper Proterozoic deposits rest on Archean gneisses and granites and are exposed in several isolated areas, the largest of which occur in the Srednii and Rybachii peninsulas (Fig. 1), Kil'din Island, and along the Terskii coast of the southern Kola Peninsula. Insignificant exposure areas are recorded in the local graben-shaped depressions in the northern and eastern coasts of the Kola Peninsula. The most complete sections of the deposits under consideration are located on the Srednii and Rybachii peninsulas and on

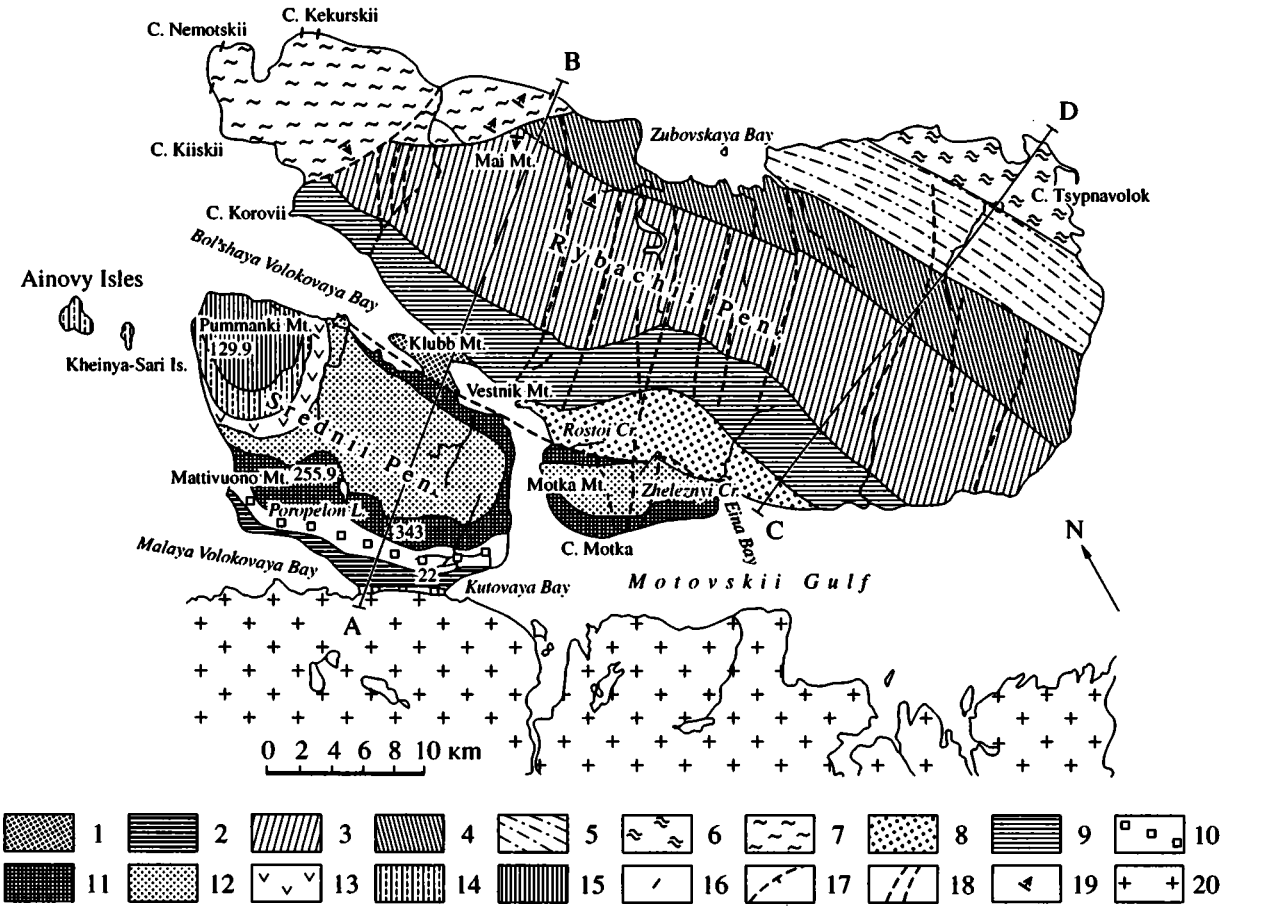

Fig. 1. Schematic geological map of the Rybachii and Srednii peninsulas, the Malyi (Kheinya-Sari) and Bol'shoi Ainovy Islands (compiled by V.V. Lyubtsov and A.A. Predovskii with regard for the data of T.P. Vronko, D.F. Agap'ev, and V.Z. Negrutsa).

Eina Group, Upper Riphean formations: (1) Motka (polymictic conglomerates, subarkoses); (2) Lonskaya (conglomerates, shales, and subgraywacke sandstones); (3) Pereval'naya (graywacke and subgraywacke sandstones)

Bargoutnyi Group, Upper Riphean formations: (4) Mai (conglomerates, gritstones, and graywacke sandstones); (5) Zubovka (siltstones and graywacke sandstones); (6) Tsypnavolok (graywacke siltstones with hydromicaceous and chloritic cement); (7) Skarbeevskii (shales and siltstones).

Kil'din Group, Upper Riphean formations: (8) Kutovaya (quartz sandstones); (9) Ierno (graywacke sandstones); (10) Palvi (subarkoses, quartzites) corresponding to the united Korovii, Bezymyannyi and Chernaya Rechka formations of Kil'din Island; (11) Poropelon (subgraywacke shales); (12) Zemlepakhtinskaya (subgraywacke–arkose, sandy); (13) Karuyarvinskaya (variegated deposits, marls, siltstones, and shales). The Poropelon, Zemlepakhtinskaya and Karuyarvinskaya formations are correlated with the Pridorozhnoe and Slantsevoe Ozero formations of Kil'din Island.

Volokovaya Group, Upper Riphean, formations: (14) Kuyakanskaya (shales, subgraywacke quartzites); (15) Pumanki (shales, subgraywackes).

Other symbols: (16) dykes; (17) faults; (18) large fracture zones; (19) bedding elements; (20) granitoids of the pre-Riphean basement.

Kil'din Island. These sections are of principal regional significance.

In 1925–1935, Polkanov suggested that rather strongly deformed deposits of the Rybachii Peninsula are younger than the deposits on both the Srednii Peninsula and Kil'din Island and are thrust over them. He reported on the same relationships of the above units in the Varanger Peninsula. In 1929, the same conclusion was made by Finnish geologist K. Vegmann, and later by Keller and Sokolov (1960; Keller *et al.,* 1963), Chumakov (1974), and Konopleva (1976). The results of "The Northern region" project on the structural zoning of the territory confirmed the Polkanov's suggestion.

The problem of time relationships between deposits of the Srednii and Rybachii peninsulas stems from the fact that they are in tectonic contact throughout the Russian territory. In contrast, in the Persfjorden area of the northeastern Varanger Peninsula, Norway, there is the Kongsfjord Formation, an analogue of deposits of the Rybachii Peninsula (Siedlecka *et al.,* 1992a, 1992b), which is distinctly overlain by the Basnaeringen Formation correlated to the lower beds of Kil'din Group. The correlation is based on the fact that the latter formation in the Persfjorden area includes interbeds and members of red sandstones characteristic of Kil'din Group and their analogues in Norway.

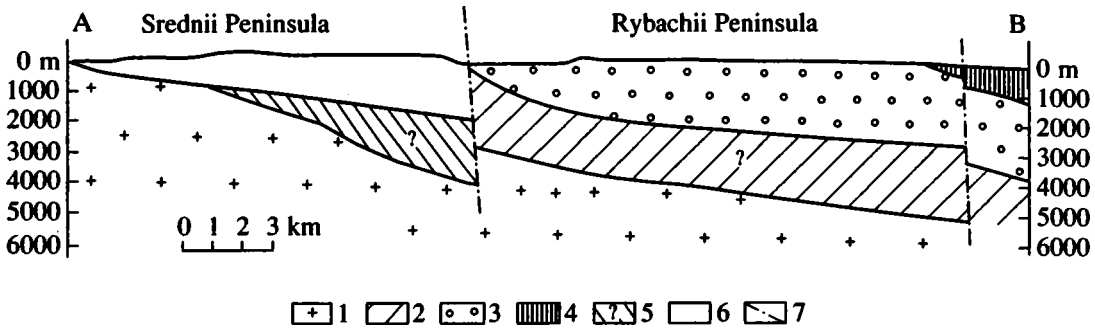

Fig. 2. Section along line AB in Fig. 1: (1) Arhean basement; (2) a hypothetical lower part of the Eina Group; (3) Eina Group; (4) Bargoutnyi Group; (5) a hypothetical lower part of Kil'din Group (or an upper part of the Bargoutnyi Group); (6) Kil'din Group; (7) large faults.

The Upper Proterozoic successions of the Srednii and Rybachii peninsulas and Kil'din Island show the northwestern strike and persistently dip northeast. They may represent the southwestern limb of a large syncline located in the Barents Sea. Recent investigations suggest the following succession of the considered deposits of the Kola Peninsula coast. The oldest Eina Group, whose lower boundary has not been determined, is overlain with apparent conformity by the Bargoutnyi Group with stratigraphically undefined upper boundary. The groups are recorded in the Rybachii Peninsula only. The younger Kil'din Group is overlain without evident unconformity but with a hiatus by the Volokovaya Group, which crowns the Upper Proterozoic succession of the Murmansk coastal area. Kil'din Group is widespread in the Srednii Peninsula, Kil'din Island, and partly in the southern Rybachii Peninsula. However, as we observed together with geologists from the Geological Survey of Norway, the exposed deposits of this group are in tectonic contact with the Eina Group (Figs. 1, 2). This inference was supported by Negrutsa, who also took part in observations and argued earlier that the Eina Group overlies Kil'din Group (Negrutsa, 1971). The Volokovaya Group occurs only in the northwestern part of the Srednii Peninsula. The southern, Terskii coast of the Kola Peninsula exposes deposits generally comparable with Kil'din Group and older beds conventionally correlated with the Bargoutnyi Group.

The Riphean–Vendian deposits of the region show facial zoning that reveals sedimentational environments in tectonic zones of a sublatitudinal extension. This zoning was first outlined by Keller and Sokolov (1960), who recognized two structural-facial elements: the outer platform zone embracing the Srednii Peninsula, the Cape Motka of the Rybachii Peninsula, and Kil'din Island, and the inner miogeosynclinal zone of the Rybachii Peninsula. Both zones were acknowledged in subsequent studies of Negrutsa (1971), Sergeeva (1973), and Lyubtsov (1975). Now the available data suggest that there were three different zones of the Late Proterozoic (Late Riphean and Vendian) sedimentation: (1) the Rybachii and (2) Kil'din zones corresponding to zones recognized by Keller and Sokolov (1960), and (3) the Terskii zone (Fig. 3).

Investigations in connection with project "The Northern Region" provided us with new data on the development of an extended disjunctive structure (Fig. 3), which separates the northeastern and southwestern structural-facial zones. This structure extends for more than 1000 km. In the Norwegian sector of the Barents Sea region, it corresponds to the Trollfiorden–Komagelva fracture zone. Further southeast, it goes between the Srednii and Rybachii peninsulas and north of Kil'din Island. In the Kanin Peninsula, the structure separates the Pae Ridge (the Kanin Stone) from the Cape Ludovatye area and continues toward the Timan Ridge. It is manifested as a system of faults and associated different-scale folds. The available data suggest that this structure of synsedimentary genesis separated the continental edge from the active continental slope in the Late Proterozoic time. The nature of this structure was elucidated in the course of long-term studies of Siedlecka (1975, 1995), who acknowledged the pioneering investigations of A.A. Polkanov, L.Ya. Kharitonov, N.S. Shatskii, B.M. Keller, and B.S. Sokolov. For the first time, this fault zone within its current contours was described by Sokolov (1964) as a northern limit of the Russian platform in the Late Precambrian time. That is why we propose to name it "the Sokolov line." The structural-facial zone to the northeast of this line is now referred to as the Barents Sea–Rybachii zone (I in Fig. 3), and the zone to the southwest (II in Fig. 3) as the Tana–Kil'din zone. The third, Terskii zone is confined to the tectonic line of the Kandalaksha Gulf and is similar to the Tana–Kil'din zone, but shows a more gentle dipping of the Riphean beds than the latter, as well as the presence of Paleozoic alkalic intrusions.

In both the Russian and Norwegian territories, the Tana–Kil'din zone displays the following principal features: (1) accumulation of siliciclastic and carbonate deposits of the uppermost Proterozoic (upper Upper Riphean and Vendian) in shallow-water and peritidal settings (the Vadso, Tanafjorden, and Vestertana groups

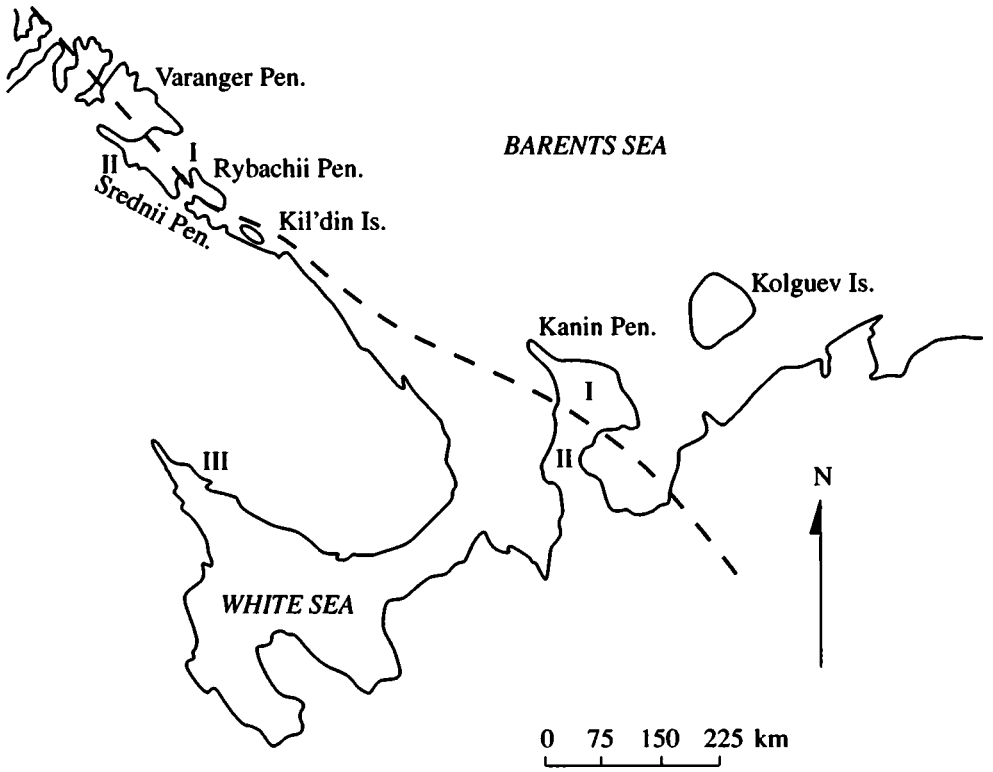

Fig. 3. The Trollfjorden–Komagelva–Rybachi–Kil'din–Kanin lineament (the Sokolov line) as a boundary of the Late Proterozoic continent in the Barents Sea region.

The lineament is designated with a stippled line. Roman figures denote structural-facial zones: (I) Barents Sea–Rybachii; (II) Tana–Kil'din; (III) Terskii.

and Kil'din and Volokovaya groups); (2) unconformable occurrence of Upper Riphean deposits above the Archean basement, which shows no signs of weathering; (3) the abundance of highly differentiated siliciclastic deposits, e.g., quartz sandstones; (4) insignificant katagenetic rock alterations; and (5) limited deformations and hydrothermal alterations of the Upper Proterozoic deposits.

Main characteristics of the Barents Sea–Rybachii zone are as follows: (1) occurrence of the Upper Riphean turbidites, including those of its lower horizons (the Kongsfjord Formation, the Eina and Bargoutnyi groups), which are extremely widespread on the Rybachii Peninsula; the lower boundary of this sequence is undetected; (2) wide areas of younger Upper Riphean and Vendian deposits (the uppermost beds of the Barents Sea Group and the Lokvikfjellet Group) on the Varanger Peninsula; (3) intense katagenetic alterations of rocks; (4) significant tectonic deformations (small folds, cleavage) and widespread development of quartz and carbonate–quartz veins of the Alpine type.

It should be emphasized that the Norwegian part of the Barents Sea–Rybachii zone (the Barents Sea subzone) is of principal significance for understanding the Upper Proterozoic stratigraphy and correlation throughout the Barents Sea region. First, as mentioned above, there are both the lower Upper Proterozoic deposits (the Kongsfjord Formation, a probable analogue of the Bargoutnyi Group of the Rybachii Peninsula) and younger horizons of the Upper Proterozoic (the Basnaeringen, Batsfjord, and Tyvjofjellet formations correlative to Kil'din Group). Second, Norwegian geologists recognized normal stratigraphic relationships between the mentioned units, indicating that the sedimentary succession in the Barents Sea subzone begins with deposits analogous to those of the Rybachii Peninsula and overlain by the correlatives of Kil'din Group; the latter, in their turn, are replaced upsection by the Vendian Lokvikfjellet Group (A. Siedlecka, 1975, 1985; A. Siedlecka and S. Siedlecka, 1967, 1971).

Thus, judging from the composition and structure of the Upper Proterozoic deposits, the transition from the Terskii zone to the Tana–Kil'din zone and farther to the Barents Sea–Rybachii zone represents a change from an intercontinental sedimentation area to the continental margin, and farther to an active continental slope. In the latter zone, tectonic activity was persistently high even in the post-Vendian time everywhere from the Tanafjorden to the Pae Ridge of the Kanin Peninsula (Predovskii, 1971; Roberts, 1972, 1993).

BRIEF CHARACTERIZATION OF UPPER PROTEROZOIC DEPOSITS ON THE KOLA PENINSULA COAST

The presented subdivision scheme of the Upper Proterozoic (Upper Riphean and Vendian) of the Kola Peninsula coast is based on all available data on adjacent territories of Russia and Norway, including the results of joint investigations with Norwegian geologists. The rock nomenclature used is based on procedure of using petrographic and petrochemical data as described by Predovskii (1980).

The Eina Group is widespread in the isthmus, connecting the Srednii and Rybachii peninsulas, and constitutes the lower part of the Riphean succession in the region (Fig. 4). The group comprises the Motka, Lonskaya, and Pereval'naya formations.

The Motka Formation (apparent thickness up to 350 m) occurs in the isthmus between the Srednii and Rybachii peninsulas and in the Cape Motka of the Rybachii Peninsula, where it is tectonically displaced near different formations of Kil'din Group. This tectonic zone represents a part of "the Sokolov line." The Motka Formation is exposed to the northeast of this zone, and Kil'din and Volokovaya groups occur southwest of it. The Motka Formation is mainly composed of undifferentiated siliciclastic deposits, mainly of alternating coarse quartz sandstones of subgraywacke and graywacke types, breccias, and polymictic predominantly granitoid boulder–pebble conglomerates replaced upsection by small-pebble conglomerates, gritstones, and quartz sandstones.

The Lonskaya Formation (700 m) consists of conglomerates, gritstones, subgraywacke and graywacke sandstones, siltstones, mudstones, and less frequent thin interbeds of clayey limestones and marls. The overlying Pereval'naya Formation (up to 2000 m) is composed of uniform coarse-grained graywacke and subgraywacke sandstones and gritstones.

The Bargoutnyi Group, comprising a single sedimentary sequence in the Rybachii Peninsula, consists of the Mai, Zubovka, and Tsypnavolok formations (Fig. 4). In addition, the so-called Skarbeevskii Formation, which is exposed in an isolated tectonic block, is commonly assigned to this group as a probable correlative of the Tsypnavolok Formation (Fig. 4).

According to Negrutsa (1971), the Mai Formation (250 m) is composed of graywacke sandstones and gritstones similar to those of the Pereval'naya Formation of the Eina Group, as well as of boulder–pebble conglomerates. The Zubovka Formation (500 m), which is of similar composition, is constituted by alternating polymictic sandstones, gritstones, and slates, the latter locally with carbonate matter. Characteristic features of this unit are the landslide folding and mud pellets. The Tsypnavolok Formation (up to 250 m) that crowns the Rybachii Peninsula section is distinct from the Zubovka Formation due to the appearance of rare carbonates, strong pyrite mineralization in dark-colored slates, and the absence of sandstones. In general, the Tsypnavolok Formation is characterized by presence of widespread graywacke sand- and siltstones in association with less common mudstones, by abundant chlorite in the matrix of sandstone and siltstone, and by low concentrations of organic matter in dark-colored siltstones and mudstones. The Skarbeevskii Formation (apparent thickness up to 500 m), correlative to the Tsypnavolok Formation, consists of polymictic sandstones, dark gray siltstones, mudstones, marls, and limestones, all intercalated with interbeds of black and dark gray shales and lenses of small- to medium-pebble conglomerates (Negrutsa, 1971).

In general, the Eina and Bargoutnyi groups of the Rybachii Peninsula demonstrate a high proportion of turbidites, which is indicative of a high water dynamic during the sediment accumulation.

Kil'din Group occurs on the Srednii Peninsula and Kil'din Island. On the southern part of the peninsula (Lyutkevich and Kharitonov, 1958; Negrutsa, 1971; Lyubtsov, 1973, 1975) it unconformably rests on the crystalline basement lacking signs of subaerial weathering. Six formations of the group are clearly traceable throughout the Srednii Peninsula (Fig. 5) (Lyubtsov *et al.,* 1989). Negrutsa (1971) described two lower formations as a single Pyaryayarvi Formation.

The basal, Kutovaya Formation (40 m) is mainly composed of quartz and feldspar sandstones, whose porous cement includes fine hydromica flakes, dolomite and glauconite grains. The overlying Ierno Formation (200 m) is rich in sandstones usually corresponding in composition to graywackes and subgraywackes. Arkoses occur in lesser amounts, and some interbeds of dark-colored hydromicaceous mudstones, siltstones, and sandstones with carbonate concretions appear at the top of the unit. Upsection, is the Palvi Formation (160 m), comprising 12 horizons of massive compact coarse-grained quartzites (from 1.0 to 6.0 m thick), which alternate with thinner beds of sandstones, silty shales, and siltstones, including thin interbeds of gritstones and small-pebble conglomerates. Quartzites display a high content of silica (85–98%) and are cemented by carbonate- and glauconite-bearing matrices.

The younger Poropelon Formation (280 m) is represented by massive sandstones with carbonate cement, and these rocks enclose interbeds and lenses of small-pebble conglomerates and siltstones. This unit is overlain with a sharp boundary by the Zemlepakhtinskaya Formation (500 m). Cross-bedded coarse sandstones of peculiar yellowish-gray coloration and a high (2–5%) content of titanium dioxide in some thin interbeds are typical of the last formation. The lower 10–15 m of the unit contain a bed of gritstone conglomerates 2.5–3.0 m thick and containing black mudstone pebbles and small phosphorite nodules. The succession is crowned with the Karuyarvinskaya Formation, exposed only in the northeastern and western coasts of the Srednii Peninsula and in the middle course of the Vykat River. The

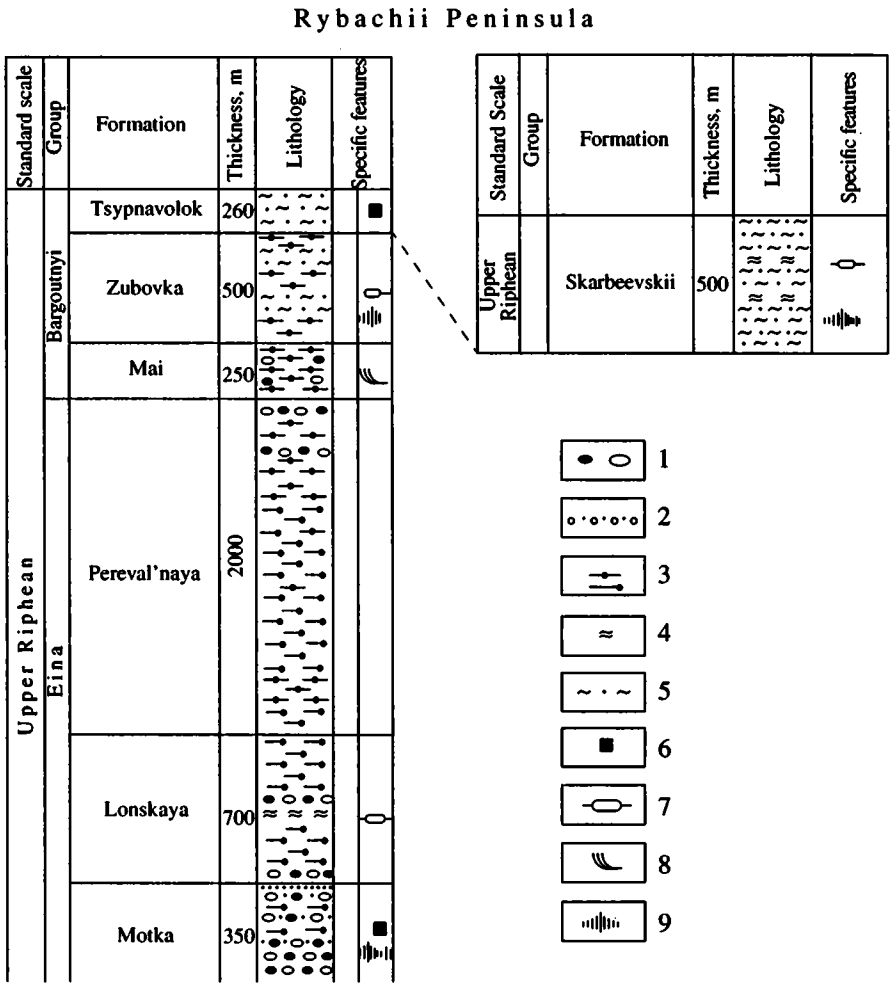

Fig. 4. Upper Proterozoic lithostratigraphy of the Rybachii Peninsula: (1) polymictic, predominantly granitoid conglomerates and gritstones; (2) subgraywackes and arkosic sandstones; (3) graywackes and silty sandstones; (4) hydromicaceous shales; (5) siltstones, predominantly of the graywacke type, with hydromica–chlorite cement; (6) syngenetic and diagenetic iron sulfides, predominantly pyrite; (7) concretions (carbonate and carbonate–clayey); (8) cross bedding; (9) landslide structure.

formation is represented by alternating red carbonate sandstones of subgraywacke type, siltstones, mudstones, and less abundant gray to variegated dolomites. An alternation of red and gray-green interbeds is characteristic. On the western Srednii Peninsula (see Fig. 1), the upper horizons of the Karuyarvinskaya Formation extend northward along the coast line for 2.5–3.0 km, and this suggests the absence of a significant angular unconformity between Kil'din and Volokovaya groups, although there is an evident hiatus.

On Kil'din Island, Kil'din Group consists of a number of local formations (Fig. 5), which can be correlated with the formations of the Srednii Peninsula (Lyubtsov *et al.*, 1978). The correlation is based on lithostratigraphic criteria, particularly on occurrence of products of strong sedimentary differentiation (quartz sandstones, carbonate-bearing and carbonate deposits) at the correlation levels, on similar alternation of beds comparable in composition, and on other features indicating similar sedimentation environments. The most reliable correlation levels consist of deposits formed under the stable tectonic regime and related strong differentiation of sedimentary material. However, facies variability of coeval deposits along and across "the Sokolov line" hampers their direct lithological correlation.

Of seven successive units of Kil'din Group, four lower formations (Ierno, Korovii, Bezymyannyi, and Chernaya Rechka) are exposed in the coastal area of the southern Kil'din Island, and three upper formations (Pesets Ozero, Pridorozhnoe, and Slantsevoe Ozero) occur in the major elevated part of the island.

Basal beds of sandy–clayey Ierno Formation were penetrated in the early 1930s by a drillhole (data of A.A. Bogdanov and P.A. Gurvich cited by Lyutkevich and Kharitonov, 1958), and its upper part is exposed. The overlying Korovii Formation (55 m) is represented by silty sandstones with lenses of glauconite sand-

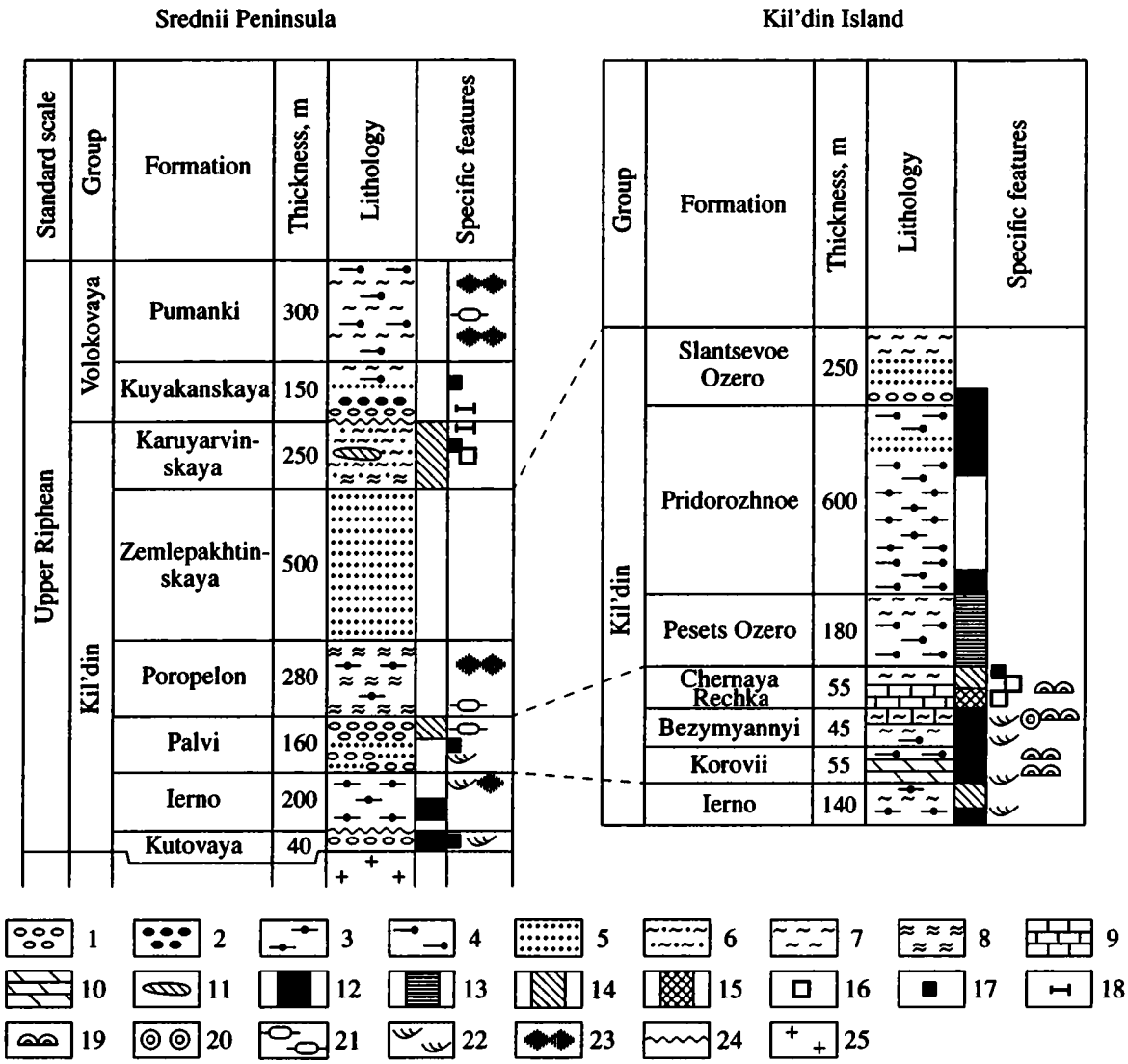

Fig. 5. Upper Proterozoic lithostratigraphic sections of the Srednii Peninsula and Kil'din Island and their correlation according to Lyubtsov and Predovskii with regard to data of Negrutsa (1971), Lyubtsov (1975), and Lyubtsov *et al.* (1978): (1) quartz sandstones and gritstones; (2) arkosic and feldspar sandstones and gritstones; (3) graywacke sandstones and silty sandstones; (4) subgraywacke sandstones; (5) subgraywacke and arkosic sandstones; (6) siltstones, predominantly subgraywacke and graywacke types with hydromica cement; (7) hydromicaceous shales (pelitic and silty); (8) hydromicaceous shales (predominantly pelitic) forming persistent horizons; (9) limestones and dolomitic limestones; (10) dolomites and calcareous dolomites; (11) lenses of carbonate and marl; (12) levels of persistent glauconite occurrence; (13) gray deposits in combination with variegated horizons and interbeds; (14) variegated deposits; (15) red deposits; (16) syngenetic hematite and other ferric oxides in sediments; (17) syngenetic and diagenetic ferric sulfides, commonly pyrite; (18) phosphate occurrences (concretions, pebbles, and matrix); (19) stromatolites; (20) oncolites; (21) concretions (carbonate and carbonate-clayey); (22) cross bedding of different types; (23) submarine landslide forms; (24) erosion surfaces or hiatus; (25) granitoids of pre-Riphean basement.

stones and interbeds of stromatolitic limestones and dolomites bearing many taxa, which characterize one of the Upper Riphean stromatolitic horizons in northwestern Russia (Krylov and Lyubtsov, 1976; Raaben and Lyubtsov, 1993; Raaben *et al.*, 1995). The Bezymyannyi Formation (45 m) above this unit contains subgraywacke siltstones and silty sandstones with interbeds of glauconite sandstones and oncolitic and stromatolitic (in the upper part) dolomites.

The boundary between the Bezymyannyi and conformably overlying Chernaya Rechka formation separates deposits different in composition and color. The Chernaya Rechka Formation (55 m) predominantly consists of red (at the base) and variegated (upsection) silty shales, siltstones, silty sandstones, and shales. Some of its shale interbeds (1–3 m thick) are rich in carcareous material ($CaCO_3$ from 5 to 30%). In addition, the formation contains thin interbeds of limestones and stromatolitic dolomites in its middle part.

Correlation scheme for Upper Proterozoic deposits in the Kola Peninsula coast and the Varanger Peninsula with due regard for structural-facial zoning

Standard scale	Rybachii–Barents Sea zone				Tana–Kil'din zone						Terskii zone
	Barents Sea subzone		Rybachii subzone		Tana subzone		Srednii subzone		Kil'din subzone		
	Group	Formation	Group	Formation	Group	Formation	Group	Formation	Group	Formation	Formation
Vendian	Lokvikfjellet	Skidnefjellet, Stordalselva, Skjaergardsneset, Styret, Sandfjorden			Vestertana	Breivika, Stappogiedde, Mortensnes, Nyborg, Smalfjord					
							Volokovaya	Pumanki, Kuyakanskaya			
Upper Riphean	Barents Sea	Tyvjofjellet			Vadso, Tanafjorden	Grasdalen, Hanglechaerro, Vagge, Gamasfjellet, Dakkovarre, Stangenes, Gronneset, Ekkeroya, Golneselva, Paddeby, Andersby, Fugleberget, Klubbnasen, Veinesbotn	Kil'din	Karuyarvinskaya, Zemlepakhtinskaya, Poropelon, Palvi, Ierno, Kutovaya		Slantsevoe Ozero, Pridorozhnoe, Pesets Ozero, Chernaya Rechka, Bezymyannyi, Korovii, Ierno	Chapoma
		Batsfjord									
		Basnaeringen									
		Kongsfjord	Bargoutnyi	Tsypnavolok (and Skarbeevskii)							Terski (conventionally)
				Zubovka							
				Mai							Turii (conventionally)
			Eina	Pereval'naya							
				Lonskaya							
				Motka							

Variegated coloration is also peculiar to the overlying Pesets Ozero Formation (180 m), represented by clayey and silty sandstones of subgraywacke and graywacke types, and also by sandstones and shales with abundant, frequently sericitic or maculose dolomitic cement. Glauconite-bearing quartz sandstones and red sandy dolomites with clayey admixtures are characteristic of the formation. Upsection, these deposits are replaced by the thick (up to 600 m) and monotonous Pridorozhnoe Formation, composed of subgraywacke, graywacke and arkosic sandstones of gray to greenish-gray color. At the base and top of the formation, these sandstones contain a noticeable glauconite admixture. Finally, the Slantsevoe Ozero Formation (250 m) that crowns Kil'din Group of Kil'din Island shows the predominance of irregularly grained subgraywacke, arkosic and feldspar–quartz sandstones. Basal planes of sandstone beds demonstrate signs of sediment reworking and concentration levels of ore minerals.

The Volokovaya Group is widespread in the northwestern Srednii Peninsula (Figs. 1, 5). In all sections examined, it rests on variegated deposits of the upper Karuyarvinskaya Formation. The group comprises the Kuyakanskaya and Pumanki formations (Negrutsa, 1971). Distinct signs of erosional disconformity occur at the base of the group.

The Kuyakanskaya Formation (150 m) is composed of coarse-grained arkosic and quartz sandstones and less frequent quartz gritstones, which are associated with interbeds of conglomerates and breccias in the lower part of the formation and with beds of subgraywacke sandstones, including silty shale interbeds, in the upper part. In the northeastern Srednii Peninsula, a sharp boundary is observable between the variegated upper part of Kil'din Group and the Kuyakanskaya Formation of the Volokovaya Group. The latter consists of gray coarse-grained quartz sandstones with lenses and interbeds of conglomerates including acute-angle rock fragments of the Karuyarvinskaya Formation, small quartz pebbles, rounded phosphorite concretions, and granitic clasts. The gravelly quartz sandstones of the Kuyakanskaya Formation fill fractures between divergent and partly displaced blocks of the variegated deposits of the Karuyarvinskaya Formation. Higher in the section, the basal horizon of the Volokovaya Group includes coarse clastic interbeds resembling olistostromes (Lyubtsov *et al.*, 1978). The Pumanki Formation (300 m), completing the succession under consideration, is represented by rhythmically alternating horizons of subgraywacke sandstones and silty shales. In general, the Volokovaya Group is distinct because of landslide deformations and discrete signs of active hydrodynamics during sedimentation.

The described Upper Riphean and Vendian deposits are largely correlative with some fragments of the same sequence, which are exposed in other Kola coast areas, such as the Ainovy Islands (the upper Kil'din and the lower Volokovaya groups) and poorly exposed coastal areas of the Srednii Peninsula and Kil'din Island (analogues of Kil'din Group).

Upper Proterozoic deposits of the southern Terskii coast of the Kola Peninsula demonstrate specific features. These are the rhythmically bedded, predominantly graywacke sandstones of the Tur'ev Formation (the Cape Tur'ev), usually assigned to the Middle Riphean rocks, the platform-type red and variegated siltstones and sandstones of the Terskii Formation (Sergeeva, 1973), arbitrarily referred to the lower Upper Riphean, and red deposits of the Chapoma Formation, correlative with Kil'din Group of the upper Upper Riphean (Konopleva and Fanderflit, 1976; Timofeev, 1969; Lyubtsov *et al.*, 1989).

AGE AND CORRELATION OF UPPER PROTEROZOIC DEPOSITS OF THE KOLA PENINSULA COAST

The available radiometric dates for deposits of the Kola Peninsula coast are represented by few K-Ar and Rb-Sr age determinations for sedimentary minerals (for review and references see Samuelsson, 1995). The K-Ar dates for glauconite from Kil'din Group vary in the range of 1059–619 Ma, whereas K-Ar and Rb-Sr dates for glauconite and illite from the Volokovaya Group vary from 900 to 670 Ma. The same determinations for the Eina and Bargoutnyi yielded age values ranging from 887 to 715 Ma and probably "rejuvenated" due to the Baikalian events in the Barents Sea–Rybachii zone. Thus, the available radiometric dates generally confirm the Late Riphean–Vendian age of the studied deposits, but they do not provide sufficient grounds for a detailed geochronology.

In this respect, the investigation results on organic-walled microfossils (Lyubtsov *et al.*, 1989; Samuelsson, 1995) are more promising. These results are consistent and indicate the Upper Riphean age of Kil'din and Volokovaya groups, and compositions of taxa from the Karuyarvinskaya Formation of the upper Kil'din Group and from the Volokovaya Group suggest that these units belong to upper horizons of this erathem. The Eina and Bargoutnyi groups also contain Late Riphean microfossils. However, they appear to occupy a lower stratigraphic position relative to Kil'din Group and thus to belong to the lower horizons of the Upper Riphean.

A comparative analysis of stromatolites from an area extending from the Porsanger in the west to the Kanin Peninsula in the east (Raaben *et al.*, 1995) revealed a number of characteristic assemblages of these biolites and supported the following correlations. The lower Lyudikovatyi Formation of the Kanin Peninsula corresponds to the lower Minka Horizon of the Upper Karatavian (the upper Upper Riphean). The upper part of Lyudikovatti Formation, the Korovii Formation of Kil'din Group, and, probably, the Porsanger dolomites correlate with the lower part of the younger

Suleimanovo Horizon of the Upper Karatavian. The Grasdalen Formation of the Tanafjorden Group corresponds to the upper part of the mentioned horizon, representing the major lower part of the Upper Karatavian. Stromatolites of the terminal Karatavian, i.e., of the highest horizons of the Upper Riphean type succession, have not been found in the described region (Raaben *et al.*, 1995).

The detailed correlation of Riphean and Vendian sequences of the Barents Sea region is the aim of future studies. The correlation between deposits of the Kola Peninsula coast and northern Norway (the Varanger Peninsula), which are based at present on joint investigations with Norwegian geologists (Siedlecka *et al.*, 1992a, 1992b), are presented in the table.

The table shows that rock associations (the Kongsfjord Formation and the Eina and Bargoutnyi groups) of tectonically active early stages of sedimentation are distinct because of widespread graywacke turbidites. According to this feature, they can be arbitrarily correlated to the Turii Formation, the lower unit of the Upper Proterozoic sequence in the Terskii zone. In addition, the above units contain no red deposits typical of younger horizons of the sequence under consideration. It is noteworthy that deposits of the early stages have not been recognized in the exposures of the Tana–Kil'din zone. The problem of their subsurface occurrence can be resolved by drilling in the northern Srednii Peninsula (Fig. 2).

The younger Late Riphean strata are more widespread. They occur throughout the Tana–Kil'din zone, are suggested to occur in the Terskii zone, and are recognized in the Barents Sea subzone of the Barents Sea–Rybachii zone. However, they are missing in the Rybachii subzone of the last zone. This part of the Upper Proterozoic sequence includes (see table) the upper half of the Barents Sea Group, the Vadso and Tanafjorden groups, and also Kil'din and partly Volokovaya groups. These deposits unconformably resting on the basement (in the Tana–Kil'din zone) reveal evidence of discontinuous sedimentation and contain highly differentiated sediments (quartz sandstones, sericite shales, and carbonates, the stromatolite type including). This association suggests a stabilized tectonic regime (Predovskii, 1980) typical of a certain developmental stage of the region. Another characteristic feature of the deposits under consideration is a widespread red and variegated coloration. The correlation between the listed units of the Tana, Srednii and Barents Sea subzones can be refined after using petrochemical and mineralogical data.

The younger, Vendian deposits occur in the Barents Sea subzone of the Barents Sea–Rybachii zone, in the Tana subzone, and partly in the Srednii subzone of the Tana–Kil'din zone (see table). These are the Lokvikfjellet and Vestertana groups and, probably, an upper part of the Volokovaya Group. Their deposits include two horizons of classic tillites in the Norwegian territory and show landslide deformations in the Russian sector (Lyubtsov *et al.*, 1989). The Vendian age has been proved by Norwegian geologists (Siedlecka and Roberts, 1992; Samuelsson, 1995) for the deposits of the Varanger Island, but the accumulation time of upper portion of the Volokovaya Group in the Srednii Peninsula is still debatable.

The presented data show that in the early Late Riphean, a long period of uplift and nondeposition was replaced (mainly the Barents Sea–Rybachii zone) by the turbidite accumulation in a continental slope setting affected by active tectonic movements. Later, in the late Late Riphean, the area of sedimentation migrated to the southwest in the Tana–Kil'din zone and the Barents Sea subzone of the Barents Sea–Rybachii zone. Sedimentation occurred under stabilized tectonic conditions, responsible for a higher sedimentary differentiation. The terminal Vendian stage of the Late Proterozoic development of the region was marked by a sharp deterioration of the climate and active relief-forming processes. As a consequence of these changes, glacial deposits (Chumakov, 1974) accumulated in the western part of the region. Later, this type of sedimentation was replaced by marine siliciclastic sedimentogenesis in the Barents Sea–Rybachii zone, which acquired specific structural patterns in response to tectonic inversion (Predovskii, 1971; Roberts, 1972, 1993).

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Parallel Biostratigraphic Scales of the Boreal Bathonian and Callovian in the North Pechora Region as a Tool for Interregional Correlations

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Abstract—Investigation results on ammonites, belemnites, and dinoflagellate cysts from Bathonian and Callovian deposits in the North Pechora region are used to elaborate the detailed stratigraphy and to correlate these deposits in the northern West Europe and Siberia. The Pechora basin, which was located in Jurassic time at the boundary between the Arctic and Boreal Atlantic paleobiogeographic provinces, incorporates faunas typical of neighboring regions and thus is an intermediate province linking Siberia with West Europe. The bed-by-bed study of sections exposed along tributaries of the Pechora River provided grounds to elaborate the Bathonian–Callovian biozonation of ammonites, belemnites, and dinoflagellate cysts. The ammonite zonation consisting of eleven units is correlated with the West European standard and East Siberian local zonations. The identified common species suggest that Pechora sections include the standard *zigzag, herveyi, calloviense, athleta,* and *lamberti* zones. Other units bearing only the arctic species are correlated with Bathonian and Callovian ammonite zones established in eastern and western Siberia. The revealed belemnite zonation includes five biostratigraphic units spanning the middle Bathonian and entire Callovian intervals. It is correlated with the regional ammonite scale and also with the East Siberian belemnite zonation. The Bathonian–Callovian succession of dinocysts assemblages is shown to include seven units ranked as zones or dinocyst beds. The dinocyst assemblages from Bathonian and Callovian deposits of the North Pechora and other regions of the Russian platform, along with similar assemblages from northern Siberia, are calibrated by comparison with dinoflagellate zonations in northwestern Europe and Arctic Canada. The mutually coordinated Bathonian–Callovian zonation of ammonite, belemnite, and dinocyst fossils is suggested to be a reliable biostratigraphic scale for interregional correlations in a vast area of northern Eurasia, Arctic regions, and America. The paper also includes description of *Cadoceras pishmae* Meledina, *C. variabile* Spath (ammonites), and *Belemnopsis* sp. ind.

Key words: *cephalopods, dinocysts, Bathonian, Callovian, zonation, North Pechora region, Russia.*

INTRODUCTION

Middle Jurassic marine basins of northwestern Europe and Siberia resided in different paleogeographic provinces: Boreal–Atlantic in the first case and Arctic in the second (Sachs *et al.*, 1971; *Paleogeografiya* ..., 1983). In western Europe and Siberia (primarily in the eastern Siberia), biostratigraphic zonations are based on different ammonite families, genera, and species. For instance, the Bajocian, Bathonian, and Callovian zonations in East Siberia are based on a succession of ammonite species from the Cardioceratidae family, whereas other families are used to substantiate the standard zonations of the same stages, and this retards stratigraphic correlations in both regions.

The problem can be solved by using the Middle Jurassic sections of the Pechora River basin (North Pechora region). The Jurassic sea of the region, situated at the boundary of the Boreal–Atlantic and Arctic paleobiogeographic provinces, was favorable for biota mixing in certain periods of the Middle Jurassic time. The corresponding stratigraphic intervals bear here either the isolated, or combined assemblages of Boreal–Atlantic and Arctic genera and species.

The main objective of this study was to elaborate the agreeing zonations, based on ammonites, belemnites, and dinoflagellate cysts, and to use them subsequently as regional standards for interregional correlation. The investigated material includes samples collected by Meledina in 1972, when she studied outcrops along the Izhma and Pizhma rivers in the Pechora basin (Fig. 1), and also the ammonite fauna recovered by M.S. Mesezhnikov and V.S. Kravets from exposures of the Pizhma, Adz'va, and Usa rivers. In addition, we examined fossils collected by V.G. Knyazev in outcrops of the first river and palynological preparations of N.K. Kulikova from borehole 176 in the Sysola area, which were kindly given to us by the All-Russia Research Institute of Geological Prospecting.

BATHONIAN AND CALLOVIAN AMMONITE ZONATIONS CORRELATED WITH WEST EUROPEAN STANDARDS

Middle Jurassic deposits in the Pechora River basin are divided into the Sysola Formation, mainly composed of sands, and the Churkina Formation of marine clay beds, intercalated with siltstone interlayers. The former partially underlies and grades into the Churkina Formation in the lateral direction (Chirva, 1989; Chirva *et al.*, 1993). The stratigraphic range of the last formation is considered as corresponding to either the Callovian (Kravets *et al.*, 1976; *Paleogeografiya* ..., 1983) or the Bathonian–Callovian (*Stratigrafiya* ..., 1976; Chirva *et al.*, 1993). According to Chirva and Yakovleva (1982), foraminifers from the lower part of the Churkina Formation are correlative with their assemblages from Bajocian–Bathonian deposits of northeastern Siberia, and thus the Bathonian beds seem to be present in this unit. The foraminiferal fauna occurs at a lower stratigraphic level than the early Callovian ammonites of *Cadoceras* genus, the guide fossils for age determination of the entire sequence of marine clay

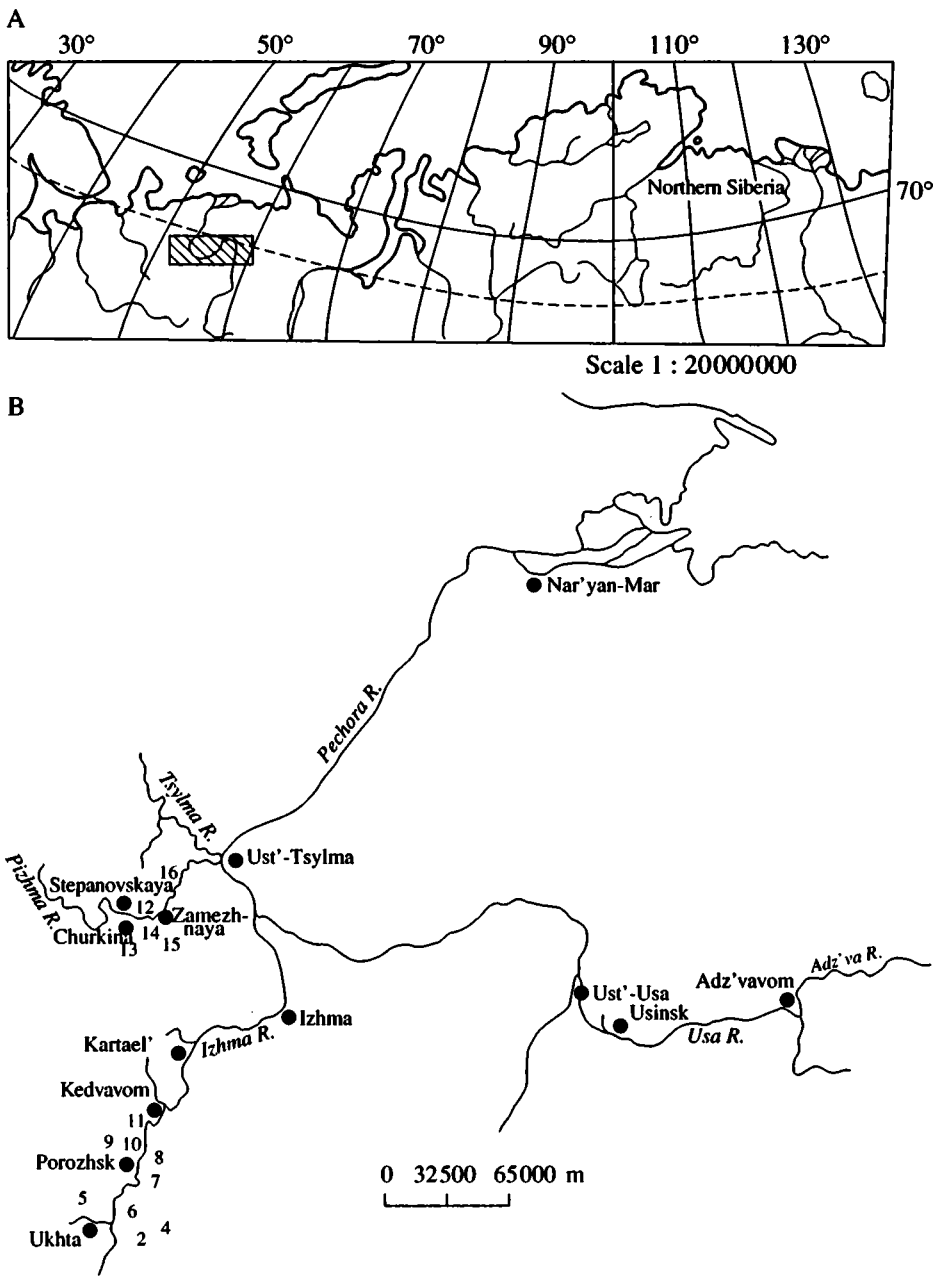

Fig. 1. Geographic position of studied area (hatched in A) and localities of Bathonian and Callovian deposits exposed along the Izhma and Pizhma rivers (numbered circles in B).

beds. Dinocyst assemblages, whose stratigraphic ranges are correlated with Bathonian dinoflagellate zones of the standard section in England, also imply that the basal interval of the Churkina Formation is Bathonian in age (Ilyina, 1991).

Zonation and successive changes in ammonite assemblages from the Churkina Formation are studied in local sections limited in thickness and exposed along the Izhma, Pizhma, Adz'va, and Usa rivers. The lower part of this formation is represented best of all in the Pizhma river valley, where clay exposures between the villages of Stepanovskaya and Churkina were formerly considered to be early Callovian in age. The basal clay beds are usually concealed under water and inaccessible for investigation. In some dry seasons, when the water level is low, there is, however, a possibility to observe horizons of carbonate concretions embedded in clay and to collect macrofossils from these rocks. The higher beds of the lower to upper Callovian interval are exposed downstream in coastal escarpments of the Pizhma River. In addition, they were studied in some outcrops of the Izhma River. The bed-by-bed description of Callovian substages in several key sections is published in several works (*Stratigrafiya* ..., 1976; Chirva and Yakovleva, 1982; Meledina, 1987), and the composite section of Middle Jurassic deposits in the Pechora River basin along with its ammonite zonation was described by Meledina and Zakharov (1996).

It has been thought for many years that Callovian deposits in the Pechora River basin comprise three substages and six zones: the local *Arcticoceras ishmae* and *Cadoceras elatmae* zones of the lower substage; the *Kosmoceras jason* and *Erymnoceras coronatum* zones of the middle Callovian; and the *Longaeviceras keyserlingi* and *Eboraciceras subordinarium* zones of its upper substage (Bodylevskii, 1963; *Stratigrafiya* ..., 1976; Meledina, 1987).

Revising the collection of Mesezhnikov few years ago, Meledina identified two early Bathonian ammonite species *Oraniceras* cf. *gyrumbilicum* Qu. and *Gonolkites* cf. *convergens* Buckm. (Meledina and Nal'nyaeva, 1991; Meledina, 1994). These species proved the early Bathonian age of basal beds of the Churkina Formation. The *Arcticoceras ishmae* Zone was also attributed to this substage (*Unifitsirovannaya stratigraficheskaya skhema* ..., 1993). In studying species of the *Arcticoceras* genus present amid ammonites from localities of the Izhma River, Meledina (1994) has concluded that they are also present in three different *Arcticoceras* zones of Siberia. She suggested the local *Arcticoceras ishmae* Zone be replaced by two units, namely by *A. ishmae* and *A. harlandi* beds corresponding, in her opinion, to the whole middle Bathonian and even to a part of the upper Bathonian substage. Like in the regional ammonite scale of the Jurassic in East Siberia, both units were placed below the *Cadoceras variabile* Beds of the upper Bathonian.

Sandstones bearing *Arcticoceras* forms are exposed at the Izhma River bank near the Dreshchanka Creek. Their beds correspond to marine clay layers barren of ammonites and known in the Bathonian section at the Churkina locality of the Pizhma River. In addition, the Churkina Formation yielded several new *Cadoceras* species.

The revised ammonite collection was used to essentially refine the subdivision scheme of Bathonian and Callovian deposits in the Pechora River basin (Table 1). Sedimentary sequences of these stages in the North Pechora region exhibit some intervals, where the West European genera and species coexist with ammonites of the Arctic province. These intervals taken together represent a kind of framework enabling correlation of regional stratigraphic schemes elaborated in this region and northwestern Europe. Using the respective reference levels we arrived at the conclusion that local units of the North Pechora region can be correlated with the following West European standard zones: *Oraniceras* and *Gonolkites* beds with the *zigzag* Zone; *Cadoceras* ex gr. *elatmae–C. falsum* and *C. pishmae* beds with the *herveyi* Zone; *C. simulans* Beds with the *koenigi* Zone; *Kepplerites* cf. *tychonis* Beds with the *calloviense* Zone; local *Longaeviceras keyserlingi* Zone with the *athleta* Zone; and *Eboraciceras subordinarium* Zone with the *lamberti* Zone.

Regional stratigraphic units between these reference levels bear only Arctic ammonite forms and can be correlated with the standard zonation less reliably. At the same time, however, they yielded some East Siberian species, and, consequently, the ammonite zonations in the last and North Pechora regions turned out to be well correlative within stratigraphic ranges of several genera and species, such as *Arcticoceras, Cadoceras variabile, C. falsum, Rondiceras, Longaeviceras, Eboraciceras,* and *Quenstedtoceras.*

The detailed substantiation of Bathonian–Callovian correlation schemes for northern Eurasia has been presented by Meledina and Zakharov (1996).

Comprehensive descriptions of ammonites, including their zonal index species, are published in a series of works (Meledina, 1977, 1987, 1994; Meledina and Zakharov, 1996) and supplemented here with new data on two species.

BATHONIAN AND CALLOVIAN BELEMNITE ZONATION IN NORTH PECHORA REGION

Many researchers have already studied the Middle Jurassic belemnites from the North Pechora region. Works of Krymgol'ts (1929), Gustomesov (1960, 1964), Saks and Nal'nyaeva (1964, 1966), Nal'nyaeva (1983), Meledina and Nal'nyaeva (1991) are devoted to a encyclopedic description of belemnites and outline their stratigraphic significance.

In the Pechora River basin, belemnites occur together with ammonites in Bathonian–Callovian deposits exposed along the Izhma and Pizhma rivers.

Table 1. Bathonian and Callovian ammonite zonations in the North Pechora and Siberian regions correlated with the European standard

Stage	Substage	Northwestern Europe	North European part of Russia (North Pechora region)	West Siberia	East Siberia	
		Standard zonation	Ammonite zones and beds	Ammonite zones and beds	Ammonite zones, subzones, and beds	
Callovian	Upper	*Quenstedtoceras lamberti*	*Eboraciceras subordinarium*	*Eboraciceras subordinarium* *Quenstedtoceras lamberti*	*Eboraciceras subordinarium*	
		Peltoceras athleta	*Longaeviceras keyserlingi*	*Longaeviceras keyserlingi*	*Longaeviceras keyserlingi*	
	Middle	*Erymnoceras coronatum*	*Rondiceras milaschevici–Kosmoceras* spp. Beds	*Kosmoceras* ex gr. *jason–Rondiceras* ex gr. *milaschevici* Beds	*Rondiceras milaschevici–Erymnoceras* Beds	
		Kosmoceras jason				
	Lower	*Sigaloceras calloviense*	*Kepplerites* cf. *tychonis* Beds	*Sigaloceras* sp.Beds	*Cadoceras emelianzevi*	
		Proplanulites koenigi	*Cadoceras simulans* Beds			
		Macrocephalites herveyi	*Cadoceras pishmae* Beds	Cadoceratinae Beds	*Cadoceras anabarense*	
			C. ex gr. *elatmae–C. falsum* Beds		*Cadoceras falsum*	
Bathonian	Upper	*Clydoniceras discus*	*Cadoceras variabile* Beds	Barren of ammonites	*Cadoceras variabile* Beds	
		Oxycerites orbis	Barren of ammonites		*Cadoceras barnstoni*	
		Procerites hodsoni	*Arcticoceras ishmae–A. harlandi* Beds		*Arcticoceras (?) cranocephaloide*	
	Middle	*Tulites subcontractus*			*Arcticoceras ishmae*	
		Procerites progracilis			*Arcticoceras harlandi*	
		Asphinctites tenuiplicatus	Barren of ammonites		*Arctocephalites* aff. *greenlandicus*	
	Lower	*Zigzagiceras zigzag*	*Oraniceras* cf. *gyrumbilicum–Gonolkites* cf. *convergens* Beds		*Arctocephalites arcticus*	*A. arcticus*
						Oxycerites jugatus

Their stratigraphic analysis revealed five units ranked as belemnite zones or beds. From the base upward through the sequence, these are the *Paramegateuthis ishmensis* Beds, *Pachyteuthis tschernyschewi* Zone, *Pachyteuthis subrediviva* Beds, *Lagonibelus beaumountianus hemisulcatus* Zone, and *Acroteuthis cuneata* Beds (Table 2).

The *Pachyteuthis tschernyschewi* Zone was originally distinguished at the Dreshchanka Creek locality and considered as corresponding to the *Arcticoceras* Beds. Moreover, this index species of belemnites was reported to occur in an upper portion of the bed sequence rich in *Arcticoceras ishmae* and *Costacadoceras* sp. forms.

P. tschernyschewi is a typical Arctic species, and outside the Pechora River basin its distribution area spans other regions of East Siberia, western Canada, and Franz Josef Land (Plate 1, no. 3). It has been recently shown that in Siberian sections the *P. tschernyschewi* Zone strictly corresponds to the *Arcticoceras ishmae* range in the ammonite zonation. We assume the same stratigraphic interval for the indicated belemnite zone in the North Pechora region. Another part of strata bearing *Arcticoceras* forms below this reference level is defined as beds of *Paramegateuthis ishmensis* (Gust.), because they are predominantly enriched in this belemnite species (Plate 1, no. 4). In East Siberia, *P. ishmensis* (Gust.) coexists with ammonites of *Arctocephalites, Cranocephalites,* and *Arcticoceras* genera; it is thus appears in the upper Bajocian and is present in the lower to upper Bathonian interval up to the *A.* (?) *cranocephaloide* Zone. In the Izhma River localities, this species is recorded in the *P. tschernyschewi* Zone. Older deposits of the region are barren of belemnite rostrums. In addition to index species, belemnite assemblages from the indicated units also include other Arctic species widespread in Siberia, such as *Pachyteuthis optima* Sachs et Naln. (ranging from upper Bajocian to lower Callovian), *Paramegateuthis timanensis* Gust. (mainly typical of Bathonian), and *Belemnopsis* = *B.* ex gr. *latesulcatus* Orb., first detected in the region.

Table 2. Bathonian and Callovian zonations in the North Pechora region

Stage	Substage	Ammonite zones and beds	Belemnite zones and beds	Dinocyst zones and beds	
Callovian	Upper	*Eboraciceras subordinarium*	*Acroteuthis cuneata* Beds	Dinocysts have not been studied	
		Longaeviceras keyserlingi		VII—*Rigaudella aemila, Pareodinia prolongata*	
	Middle	*Rondiceras milaschevici–Kosmoceras* spp. Beds	*Lagonibelus beaumountianus hemisulcatus*	VI—*Chytroeisphaeridia cerastes–Netrelytron (Kalyptea) stegastum* Beds	
	Lower	*Kepplerites* cf. *tychonis* Beds	*Pachyteuthis subrediviva* Beds	V—*Lagenadinium callovianum, Chlamydophorella* sp.	
		Cadoceras simulans Beds			
		Cadoceras pishmae Beds		IV—*Fromea tornatilis Sentusidinium* spp.	b—*Paragonyaulacysta retifragmata, P. calloviense*
		Cadoceras ex gr. *elatmae–C. falsum* Beds			a—*Fromea tornatilis*
Bathonian	Upper	*Cadoceras variabile* Beds		III—*Kalyptea diceras*	
		Barren of ammonites	Barren of belemnites	II—*Dichadogonyaulax (Ctenidodinium) sellwoodii, Protobatioladinium* sp.	
	Middle	*Arcticoceras ishmae–A. harlandi* Beds	*Pachyteuthis tschernyschewi*		
			Paramegateuthis ishmensis Beds		
	Lower	Barren of ammonites	Barren of belemnites		
		Oraniceras–Gonolkites Beds		I—*Evansia evittii–Echarisphaeridia (Batiacasphaera) pocockii* Beds	

Genus *Paramegateuthis,* occurring outside the North Pechora region in East Siberia and British Columbia (Jeletzky, 1974), has been considered by Saks and Nal'nyaeva as an Arctic taxon. Nevertheless, it was also reported from the central and western Balkan Mountains of the Tethyan province (Stoyanova-Vergilova, 1982, 1990), and this suggests the southern origin of the genus.

The occurrence of *Belemnopsis* forms in the *P. tschernyschewi* Zone is quite remarkable. In the Middle Jurassic time, this genus was widespread in both hemispheres of the Earth, especially in the lower latitudes (Andes, New Zealand, Bulgaria, and elsewhere). According to Challinor *et al.,* (1992, Fig. 23.2), it was also characteristic of the Boreal–Atlantic sector (England and northern France), and the localities of *Belemnopsis* species now known in the Izhma River basin displace eastward the northern boundary of distribution area of this thermophilic taxon.

Thus, belemnites from the *Pachyteuthis tschernyschewi* Zone and underlying *Paramegateuthis ishmensis* Beds are mostly represented by Arctic taxa. In the first unit, however, they coexist with species of southern origin.

The indicated cryophilic forms offer a possibility to trace the *P. tschernyschewi* Zone throughout the Arctic paleobiogeographic province, East Siberia inclusive. The underlying *Paramegateuthis ishmensis* Beds are assuredly correlative with *Cylindroteuthis confessa* Beds of East Siberia (Meledina *et al.,* 1991), where they also bear *P. ishmensis* Gust and occupy the same stratigraphic position relative to the *P. tschernyschewi* Zone.

The *Pachyteuthis subrediviva* Beds are confined to deposits bearing diverse ammonites of the genus *Cadoceras.* In localities of the Pizhma River, this index species spans the stratigraphic interval from the *Cadoceras* ex gr. *elatmae* Beds (presumably together with some strata below them) to the *Kepplerites* cf. *tychonis* Beds of the lower Callovian, whereas, in the Izhma River basin, it occurs even higher in the middle Callovian. In central areas of the Russian platform, this species is characteristic of the *Cadoceras elatmae* Zone. It was also reported from the Clayton Bay Clay of England, dated back to the lower Callovian. In the Bol'shoi Begichev Island of East Siberia, the same taxon was detected in the *C. anabarense* Zone of the lower Callovian.

In addition to *P. subrediviva,* the zonal assemblage in question includes *Pachyteuthis optima* Sachs et Naln., *P. bodylevskii* Sachs et Naln., and *Cylindroteuthis subextensa* (Nik.). All these species are typical of middle Bathonian–lower Callovian deposits in East Siberia, where they have been detected at various levels

in the interval of *Arcticoceras* (?) *cranocephaloide* to *Cadoceras emelianzevi* zones and considered as characteristic of the *Pachyteuthis* spp. Beds. In Siberia, the last unit remains undivided.

The *Lagonibelus beaumountianus hemisulcatus* Zone spans the middle Callovian. In addition to index species, it includes *Cylindroteuthis subextensa* Nik. and *Acroteuthis krimholzi* Gust., widespread in the North Pechora region and associated with the less abundant *Pachyteuthis subrediviva* (Lem.). Sachs and Nal'nyaeva considered *L. beaumountianus hemisulcatus* Sachs et Naln. as a local Siberian subspecies of *L. beaumountianus* (Orb.) widespread in western and eastern Europe, where it is especially characteristic of the middle Callovian in England and France (Gustomesov, 1964; Krymgol'ts, 1929).

In the Nizhnii Novgorod, Kostroma, and Orenburg regions, *L. beaumountianus* (Orb.) was detected in the *Kosmoceras jason* and *Erymnoceras coronatum* zones (Kamysheva-Elpat'evskaya *et al.*, 1959; Nal'nyaeva, 1986). It was also reported as occasionally occurring in the *Peltoceras athleta* Zone.

L. beaumountianus forms can be considered as index species of the middle Callovian in Europe. In the north of European Russia, they are represented by local subspecies also characteristic of this substage.

In East Siberia, where the middle Callovian is studied in the Anabar region and Bol'shoi Begichev Island, belemnites are rare and insufficiently diverse. Accordingly, their taxonomic composition is poorly known.

The *Acroteuthis cuneata* Beds, corresponding to upper Callovian *Longaeviceras keyserlingi* and *Eboraciceras subordinarium* zones in localities at the Izhma and Pizhma rivers, bear few of the species *A. cuneata* Gust., *Lagonibelus beaumountianus* (Orb.), and *Cylindroteuthis* sp. The index species of the beds is also reported to be from northeastern Siberia. Outcrops at the Izhma River show that the late Callovian genus *Eboraciceras* and early Oxfordian genera *Cardioceras* and *Pavloviceras* often coexist with rostrums of *Pachyteuthis*, which cannot be identified at the species rank. The established belemnite zonation and its correlation with the regional ammonite zonation in the Pechora River basin are shown in Table 2. A brief description of genus *Belemnopsis* is presented at the end of the paper.

BATHONIAN AND CALLOVIAN DINOCYST ZONATION IN THE NORTH PECHORA REGION

Dinocysts represent a promising group of microfossils widely used in the interregional correlation of marine deposits. Their zonation in Bathonian, Callovian, and Oxfordian deposits of the Russian platform is based on distribution ranges and taxonomic composition of dinocyst assemblages from the Moscow syneclise, Voronezh anteclise, and Pechora basin (Ilyina, 1991a, 1991b; *Unifitsirovannaya stratigraficheskaya skhema* ..., 1993). At present, however, this zonation is in need of revision, because some dinozones have been more precisely correlated with the Bathonian–Callovian ammonite zonation, recently refined in the Russian platform (Meledina, 1994; Meledina and Zakharov, 1996). In addition, taxonomic characteristics of zonal dinocyst assemblages are now better studied in the North Pechora region due to considerable progress in the taxonomy and diagnosis of dinocyst species, which is a result of coordinated work of V.I. Ilyina, V.A. Fedorova (St. Petersburg), and J.B. Riding (England) on the International project "Mesozoic palynozonations in the Northern Hemisphere."

In the North Pechora region, Middle Jurassic dinocysts were first discovered by T.F. Vozzhennikova, who studied core samples from wells drilled in the region of the Yarenga and Bol'shaya Vizinga rivers, and these data were used to prove the marine origin of Callovian deposits (Teodorovich and Vozzhennikova, 1971). Ilyina (1991) collected these microfossils from natural exposures at the Pizhma and Izhma rives and studied them in detail. She investigated the lower part of the Churkina Formation near the village of Stepanovskaya at the left-hand bank of the Pizhma River (Fig. 2) and coeval beds exposed downstream near the village of Churkina, where they include concretions with Bathonian ammonites.

Basal beds of the studied sequence (Exposure 12, Sample 12/7) yielded the dinocyst assemblage of *Evansia evittii* (Pocock) Jans., *Escharisphaerida pocockii* (Sarj.) Erkmen et Sarj. (*Batiacasphaera* sp. in the opinion of J.B. Riding), *Pareodinina* sp., and ? *Lithodinia valensii* (Sarj.) Fenton. These species, inherited from the Bajocian, are widespread in the lower Bathonian, and their assemblage corresponds to that from the *Evansia evittii, Escharisphaerida pocockii*, and *Dichadogonyaulax* sp. beds spanning the uppermost Bajocian and lowermost Bathonian interval in the Russian platform (Ilyina, 1991a; *Unifitsirovannaya stratigraficheskaya skhema* ..., 1993). Near the village of Churkina, clay layers with this dinocyst assemblage approximately correspond to the lower Bathonian *Oraniceras–Gonolkites* Beds. A spore–pollen assemblage from the studied rock samples also suggests the early Bathonian age of deposits under consideration; however, we cannot exclude a possibility that they partially comprise the uppermost Bajocian beds.

Next, the dinocyst assemblage is recovered from a clay bed (Sample 12/3) intercalated with sands representing main rock types in the upper part of the section. It consists of *Dichadogonyaulax sellwoodii* Sarj. and *Protobatioladinium* sp., associated with less abundant *Nannoceratopsis pellucida* (Defl.) Evitt, *Pareodinia* spp., *Dissiliodinium* sp., and rare *Fromea tornatilis* (Drugg) Lentin et Williams and *Sirmiodinium grossii* (Alberti), Warren widespread in higher Jurassic horizons. This assemblage is characteristic of the *Dichadogonyaulax sellwoodii–Protobatioladinium* Zone of the Bathonian (excluding its basal and topmost beds), known in the

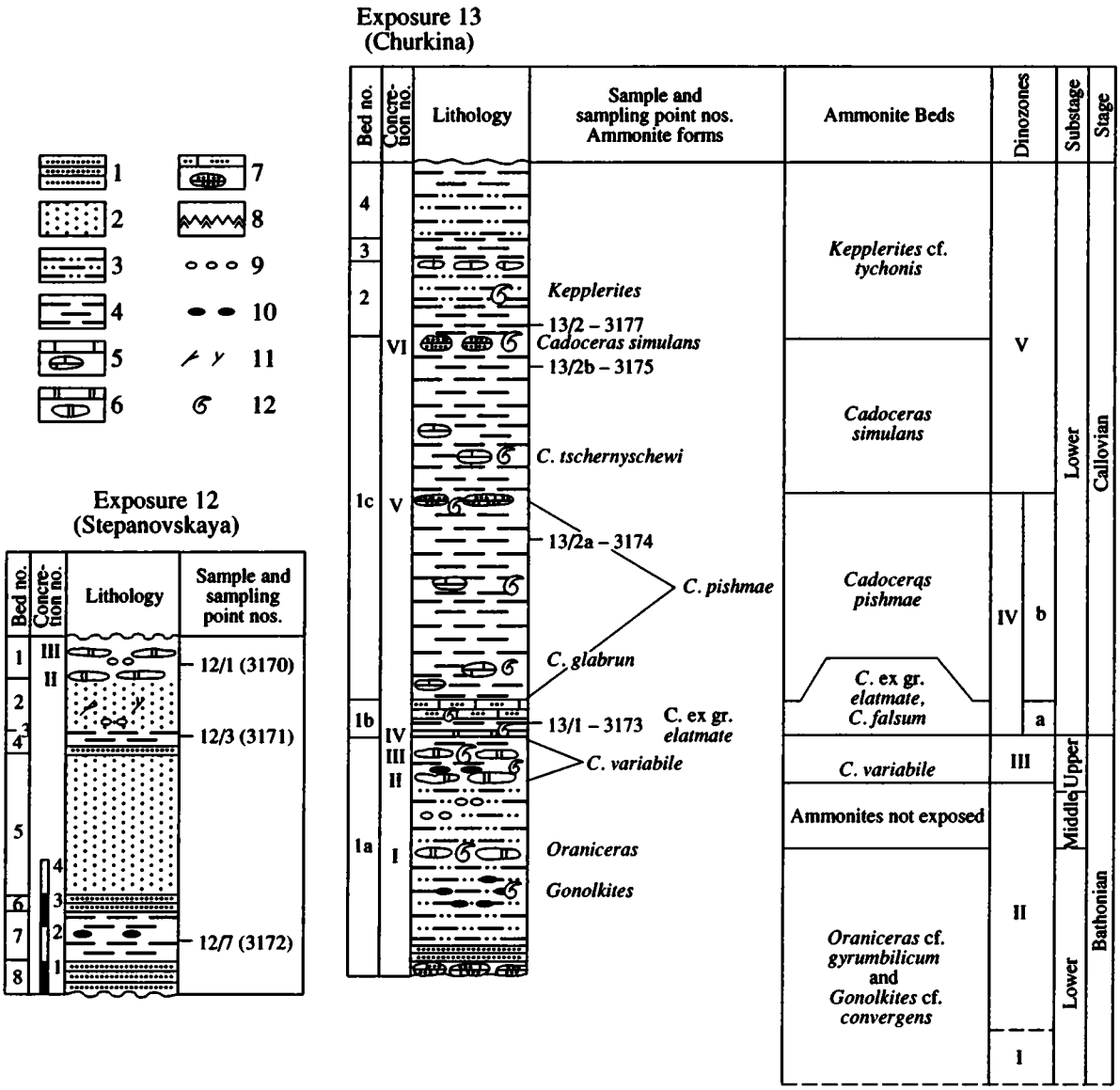

Fig. 2. Stratigraphic columns and subdivision scheme of Bathonian and Callovian deposits in the Pizhma River basin based on ammonites and dinocysts: (1) sandstone; (2) sand; (3) siltstone; (4) clay; (5) limestone beds and concretions; (6) siderite beds and concretions; (7) sandy limestone; (8) limestone with cone-in-cone structure; (9) gravel and pebbles; (10) pyrite concretions; (11) plant remains; (12) ammonite shells. Layers 1a–1c in Exposure 13 crop out in the valley of the Pizhma River, usually below the water level.

Moscow syneclise, Voronezh anteclise, and Pechora basin (Ilyina, 1991a).

In the Pizhma River basin, this dinozone presumably corresponds to the greater upper portion of the *Oraniceras–Gonolkites* Beds extending upward to the base of the *Cadoceras variabile* Beds. Its lower boundary seems to be inside the lower Bathonian, because in the Voronezh anteclise *Dichadogonyaulax sellwoodii* and *Protobatioladinium* sp. are recorded in deposits bearing the early Bathonian ammonites *Pseudocosmoceras michalskii* Boriss. and *Parkinsonia* cf. *pseudoferruginea* Nikol. (*Unifitsirovannaya stratigraficheskaya skhema ...*, 1993).

The dinocyst assemblage in question is also distinguished in the Sysola region (Well 176, depth interval of 195.2–205.0 m), where it is characteristic of the *Arcticoceras* Beds of the Churkina Formation. In addition to index species, it includes *Energlinia accolaris* (Dodek.) Sarj. and *Tubotuberella dangeardii* subsp. *primitiva* Sarj., typical of the upper half of the Bathonian in England (Sarjeant, 1982; Fensome, 1981). Specimens of *Rinchodiniopsis cladophora* (Defl.) Below, rarely occurring in the assemblage, are also known in Arctic Canada, where they are typical of dinozone correlative with the *Arcticoceras ishmae* Zone (Davies, 1983). The taxonomic composition of studied dinocysts and their position in the sedimentary sequence of the Pizhma River basin show, as well, that they span the interval of *Arcticoceras* Beds. In this and other studied regions of

the Russian platform, the upper limit of *Dichadogonyaulax sellwoodii–Protobatioladinium* Zone corresponds to the first occurrence level of *Kalyptea diceras* Cooks. et Eisen. (Ilyina, 1991a).

The considered dinozone with dominant *Dichadogonyaulax (Ctenidodinium) sellwoodii* and related species approximately corresponds to subzones "a" and "b" of the *Ctenidodinium sellwoodii–C. combazi* Zone, spanning the greater part of the Bathonian Stage (without the *discus* Zone) in England (Riding and Thomas, 1992).

In the upper part of Exposure 12 at the Pizhma River, there are clay beds bearing the ammonite species *Cadoceras variabile* Spath., which has been detected in another outcrop located downstream near the village of Churkina (Fig. 2). The dinocyst assemblage from these beds (Sample 12/1) includes *Kalyptea diceras,* associated with frequent *Nannoceratopsis pellucida, Pareodinia* spp., *Energlinia accolaris*, rare *Sirmiodinium grossii, Fromea tornatilis, Dissiliodinium* sp., and occasional *Protobatioladinium* sp. The assemblage simultaneously marks the first occurrence level of *Chytroeisphaeridia hyalina* (Raynaud) Lentin et Williams. It is typical of the *Kalyptea diceras* Zone located at the top of Moksha Formation in the Moscow syneclise and recognized in other areas of the Russian platform. In all the sections studied, this unit underlies the lowermost dinozone of the Callovian. Formerly, it was considered as corresponding to the stratigraphic interval of uppermost Bathonian and lowermost Callovian (Ilyina, 1991a). In the Pechora basin, the *Kalyptea diceras* Zone is identical in range to the *Cadoceras variabile* Beds, and accordingly we attribute to the upper Bathonian. The same stratigraphic range of this unit is now accepted in other areas of the Russian platform.

In the North Pechora region, Bathonian dinocyst assemblages coexist with foraminifers occurring at the correlation level of *Riadella sibirica* assemblage characteristic of Bathonian deposits in many regions of the Arctic paleobiogeographic province (Yakovleva, 1993). In contrast, coeval deposits of North Siberia are almost barren of dinocysts, except for occasional *Pareodinia* sp., and ? *Lithodinia* sp. Microphytoplankton is represented here only by acritarchs (*Leiofusa* spp.) rare *Veryhachium* species, and prasinophytes (*Pterospermella*).

The lower Callovian succession of dinocyst assemblages was studied in the Churkina Formation section near the village of Churkina at the right-hand bank of the Pizhma River (Fig. 2, Exposure 13).

The lower part of this section (samples 13/1 and 13/2a) is attributed to *Fromea tornatilis–Sentusidinium rioultii* dinozone recognized in the lower Callovian ammonite zone of *Cadoceras elatmae* in the Moscow syneclise (Ilyina, 1991a). In the North Pechora region, this dinozone corresponds to the *Cadoceras* ex gr. *elatmae–C. falsum* and *Cadoceras pishmae* beds, which are correlated with the standard *Macrocephalites herveyi* Zone of Europe.

Distribution patterns of dinocyst taxa in the Churkina Formation allow us to subdivide the unit in question into two subzones. The lower subzone "a" begins above the last occurrence level of *Kalyptea diceras* and includes dominant specimens of *Fromea tornatilis* and less abundant *Nannoceratopsis pellucida, Sirmiodinium grossii, Chytroeisphaeridia hyalina,* and *Pareodinia ceratophora* (Defl.) Gocht, associated with rare cysts of *Chlamydophorella* sp., *Lithodinia* sp., *Gonyaulacysta eisenackii* (Defl.) Evitt, *G. jurassica* subsp. *adecta* (Defl.) Sarj., *Rinchodiniopsis cladophora, Endoscrinium* sp., *Batiacasphaera* sp., *Tubotuberella dangeardii* Sarj., *Sirmiodiniopsis orbis* Drugg., and *Sentusidinium* spp. This subzone is approximately compatible with the *Cadoceras* ex gr. *elatmae–C. falsum* Beds.

Dinocysts from the upper subzone "b" are similar in general composition to their species in subzone "a"; however, *Fromea tornatilis* and *Sirmiodinium grossii* are less abundant here and coexist with newly appeared and plentiful Arctic species *Paragonyaulacysta retifragmata* Dorh. et Davies and *P. calloviense* (Johns. et Hills) Dorh. et Davies. This unit is concurrent with the lower Callovian *Cadoceras pishmae* Beds.

The coexistence of Boreal-Atlantic and Arctic dinocysts typical of the lower Callovian dinozone in the North Pechora region is a factor favorable for interregional correlations. For instance, the peak abundance of *Fromea tornatilis* enables correlation of the lower Callovian zones in the Pechora basin, Moscow syneclise, and Ulyanovsk–Saratov depression (Ilyina, 1991a). An assemblage including this form is also characteristic of the lower Callovian shelf deposits in the Barents Sea (Kozlova *et al.,* 1994). The monodominant *Fromea tornatilis* assemblage is reported as well from the coeval Cadoceratinae Beds known in the Shaim area of West Siberia. Moreover, rare specimens of this dinocyst form are detected in the lower Callovian *Cadoceras falsum* and *C. anabarense* zones at the western coast of the Gulf of Anabar.

Arctic species of the *Paragonyaulacysta* genus are significant for the correlation of the lower Callovian *Cadoceras pishmae* Beds distinguished in Pechora basin with the East Siberian *C. anabarense* Zone, where an assemblage demonstrating peak abundance of *Paragonyaulacysta retifragmata* is characteristic of the Gulf of Anabar area. The same species are useful for correlating the lower Callovian deposits in other regions of the Arctic paleobiogeographic province, e.g., in Arctic Canada and East Greenland, because the index species of *Lacrimodinium warrenii* Zone of the last region (Smelror, 1988) is almost identical to *Paragonyaulacysta retifragmata.*

An assemblage occurring higher in the section (Exposure 13, samples 13/2b and 13/2c) includes many species typical of the previous dinozone, although they

are less abundant here. It is distinct due to the appearance of *Lagenadinium callovianum* Piel. in association with *Chlamydophorella* sp., *Cleistosphaeridium varrispinosum* (Sarj.) Wollam et Riding, *Dingodinium* sp., *Endoscrinium* cf. *galeritum* (Defl.) Vozzen, and rare *Mendicodinium groenlandicum* (Pocock et Sarj.) Davey and *Pareodinia prolongata* Sarj. Three first taxa from this assemblage suggest that it is correlative with dinoflagellates occurring in the *Lagenadinium callovianum–Chlamydophorella* Beds, compatible with two upper zones of the lower Callovian in the Voronezh anteclise (Ilyina, 1991a). In the Pechora basin, however, index species of these beds are much less abundant than in the stratotype section and coexist with Arctic *Paragonyaulacysta* forms and rare *Crussolia perireticulata* Smelror et Arhus. In the Pizhma River basin, the stratigraphic range of the beds under consideration seems to be concurrent with *Cadoceras simulans* and *Kepplerites* cf. *tychonis* ammonite beds of the lower Callovian.

Middle and upper Callovian beds with ammonites were studied in natural outcrops of the Churkina Formation at the Izhma River (Fig. 3). Silty clay with sandstone interlayers is exposed here at the left-hand side of the river near the mouth of the Ty-Yu Creek (Exposure 9, beds 1, 2). The rocks enclose concretions with ammonites *Rondiceras milaschevici* (Nik.), and two of their samples (9/1a and 9/1b) yielded the dinocyst assemblage of abundant *Chytroeisphaeridia cerastes, Ch. hyalina, Chlamydophorella* sp., *Cleistosphaeridium varrispinosum, Fromea tornatilis, Lagenadinium callovianum, Rinchodiniopsis cladophora, Pareodinia prolongata,* and less frequent *Gonyaulacysta* spp. The assemblage is distinct because of presence of rare specimens of *Netrelytron (Kalyptea) stegastum* Sarj. It also includes *Sentusidinium pilosum* (Ehrenb.) Sarj. et Stover discovered for the first time in the region and *Lithodinia planoseptata* Riding encountered only here.

Deposits bearing the above assemblage are classified as *Chytroeisphaeridia cerastes–Netrelytron stegastum* Beds (Ilyina, 1991a). In addition to the presence of these index species, they show the maximum abundance of *Chytroeisphaeridia hyalina* and *Lagenadinium callovianum.* The base of the beds marks the first occurrence level of *Sentusidinium pilosum*. The unit spans the stratigraphic interval of the middle Callovian. Concurrent beds with a higher abundance of *Cleistosphaeridium varrispinosum* and *Lagenadinium callovianum* were also distinguished in the clay sequence with *Rondiceras milaschevici* of the Malinovyi Gorge locality situated in the Volga River basin near Saratov (Ilyina, 1991a).

Chytroeisphaeridia cerastes and *Netrelytron stegastum* are characteristic zonal species of the middle Callovian in northwestern Europe. The last of them is an index species of dinocyst zone equivalent in stratigraphic range to the *Erymnoceras coronatum* Zone of the middle Callovian ammonite zonation (Riley and Fenton, 1982).

Sandstone beds bearing *Rondiceras milaschevici* and occurring in the upper part of the section in question seem to be overlain above an insignificant hiatus by the clay bed (Exposure 9, Bed 3) formerly referred by Meledina (1987) to the middle Callovian. Near the village of Porozhskoe (Exposure 7), similar deposits lie below the concretion horizon with late Callovian ammonites *Longaeviceras keyserlingi* (Sok.). Samples from both exposures (9/1c, 7/1, and 7/2) yielded the dinocyst assemblage tightly related with the previous one, but depleted in specimens of *Chytroeisphaeridia cerastes.* Some dinocysts, such as *Chytroeisphaeridia hyalina, Cleistosphaeridium varrispinosum,* and *Lagenadinium callovianum,* do not cross the upper stratigraphic limit of this assemblage. In all the samples, the assemblage includes *Pareodinia prolongata, Mendicodinium groenlandicum, Endoscrinium* sp., *Rinchodiniopsis cladophora, Sentusidinium* spp., and occasional *Tubotuberella apatela* (Cooks. et Eisen.) Sarj. Species *Fromea tornatilis* and *Nannoceratopsis pellucida* are rather frequent at the base (Exposure 9, Sample 9/1c) of the clay bed. Another group of *Gonyaulacysta jurassica* subsp. *adecta* var. *longicornis* (Defl.) Sarj., *Stephanelytron redcliffense* (Sarj.) Stover *et al.*, and *S. (Lagenadinium) scarburghense* Sarj., together with single *Rigaudella aemula* (Defl.) Below and *Wanaea* cf. *digitata* (Cooks. et Eisen.) Wollam, represent taxa first appearing at this level.

Thus, many taxa of this last assemblage are inherited from the middle Callovian, and some of them disappear in this stratigraphic interval. At the same time, the assemblage demonstrates important changes in the taxonomic composition of dinocysts and includes several new species characteristic of the late Callovian. A case in point, judging from published data, is *Gonyaulacysta jurassica longicornis,* occurring in the late Callovian and dominant in the Oxfordian of Arctic Canada and northwestern Europe (Davies, 1983; Riding, 1984; Smelror and Below, 1992). *Stephanelytron scarburghense* and *Rigaudella aemula* are also characteristic species of the late Callovian (Riley and Fenton, 1982; Ilyina, 1991a). In Svalbard, the last species occurs at the level of *Peltoceras athleta* Zone (Smelror, 1988), and in Australia, it is recorded beginning from the base of the upper Callovian (Helby *et al.*, 1987).

Several taxa, such as *Rigaudella aemula, Pareodinia prolongata, Chytroeisphaeridia hyalina,* and *Gonyaulacysta jurassica longicornis,* suggest that the considered assemblage can be correlated with that of the upper Callovian *Rigaudella aemula–Pareodinia prolongata* dinozone in the Moscow syneclise (Ilyina, 1991a).

Accordingly, the clay bed studied in the Izhma River basin appears to be of the early late Callovian age and corresponds to the *Rigaudella aemula–Pareodinia prolongata* Zone. One should remember therewith that the *Longaeviceras keyserlingi* Zone is lesser in stratigraphic range than the *Peltoceras athleta* Zone and cor-

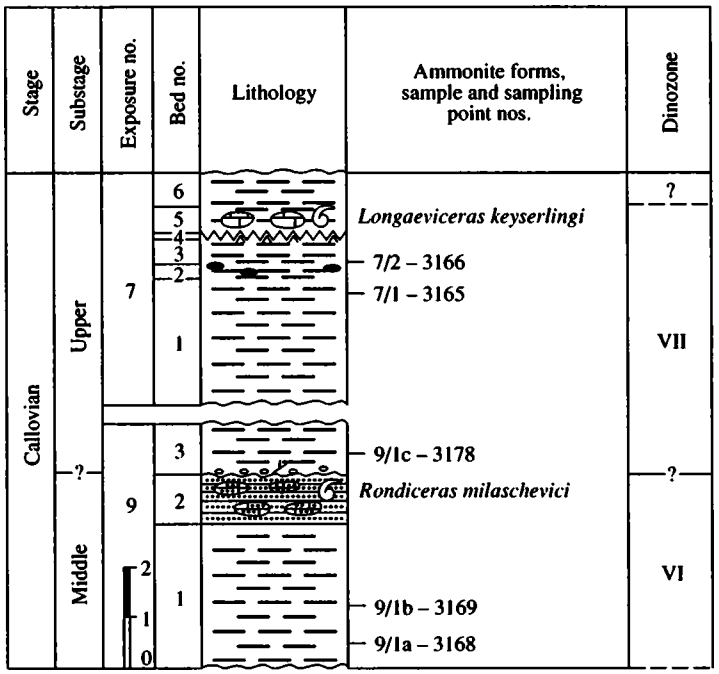

Fig. 3. Stratigraphic column and subdivision scheme of middle to upper Callovian deposits in the Izhma River basin (symbols as in Fig. 2).

responds only to two upper subzones of the latter (Meledina, 1994). Taking this into account, we assume that an interval of the Izhma River section between the *Rondiceras* and *Longaeviceras* beds, which is barren of ammonites, may correspond to the lower part of the upper Callovian.

Thus, the Bathonian–Callovian deposits in the North Pechora region reveal a succession of seven biostratigraphic units classified as dinozones and dinocyst beds, whose stratigraphic ranges are calibrated on the basis of ammonite zonation.

CONCLUSION

Ammonite, belemnite, and dinocyst zonations, which are distinguished in the Bathonian and Callovian stages of the North Pechora region, are correlated to one another (Table 2). The regional ammonite scale is correlated with the standard West European ammonite zonation on the basis of reference stratigraphic intervals enclosing the Boreal–Atlantic genera and species. Similar reference levels are also characteristic of belemnite and dinocyst zonations.

With due regard for the location of the Jurassic Pechora basin at the boundary between the Boreal–Atlantic and Arctic paleobiogeographic provinces, we believe that the generalized biostratigraphic zonation for this region can be used in practice for correlating deposits in the vast northern Eurasia and in the Boreal province as a whole.

ACKNOWLEDGMENTS

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APPENDIX

Description of ammonites

Cadoceras (Paracadoceras) pishmae Meledina, 1991
Plate nos. 1a and 1b

Cadoceras (Paracadoceras) elatmae (Meledina, 1977, p. 70; Plate 13; Plate 14, no. 1; Plate 15, no. 2; Plate 17, no. 1).

Cadoceras (Paracadoceras) pishmae (Meledina and Nal'nyaeva, 1991, p. 74; Meledina, 1994, p. 85; Plate XI, no. 2; Plate XII, no. 1; Plate XIII, no. 1; Plate XIV, nos. 1a, 1b; Plate XV, no. 1)

Holotype no. 489-209 (Meledina, 1977: Plate 13; Plate 14, no. 1; Plate 15, no. 2; Plate 17, no. 1; Meledina, 1994, Plate XIII, no. 1; Plate XIV nos. 1a, 1b; Plate XV, no. 1)

Material: three large casts with variably preserved living chambers; one of the casts is deformed.

Description. Shells are large inflated cadicones with low, moderately convolute whorls, which are rounded-trapezoid in cross sections and have the width-to-height ratios from 1.6 to 1.9. In inner whorls, the umbilical wall is gently sloping, inflection is rounded, and umbilicus itself is infundibular in form. In the course of evo-

Measured parameters of the shell (mm and percentage relative to diameter in parentheses)

Specimen no.*	Diameter, *D*	Height of whorl, *H*	Thickness of whorl, *T*	Diameter of umbilicus, *U*	Diameter of umbilical hyponome, *UH*	*T/H* ratio	Number of tubercles
489–209	127	49(37.8)	79(62.2)	46(36.2)	61(48.0)	1.63	
	102	36(35.2)	69(67.6)	37(36.2)	54(52.9)	1.90	26
489–263	103	36(34.9)	70(67.9)	37(35.9)	50(48.4)	1.90	28

* Numbers of specimens stored at the Central Siberian Geological Museum, Novosibirsk.

lution, umbilical walls grow higher and become steeply inclined, inflections turn out to be angular, and the dimensions of the umbilical hyponome increase to 1/2 of the shell diameter. The living chamber of the holotype is as long as 3/4 of the whorl. Its simple aperture is preceded by a shallow and wide inflection, and the ventral lip is slightly advanced.

Umbilical hyponomes of inner whorls display rare and narrow costae branching from the umbilical seam and slightly curved backward. Costae are inclined up ahead and elevated in the umbilical inflection in a form of elongated tubercles. Incipient points of costae gradually displace their position upward, and, when the shell is 60 mm in diameter, umbilical walls turn out to be smooth. Tubercles in the inflection exist for a longer period, although they gradually blur, as well, losing their relief. The rest of the shell is even.

Comparison. As compared to *C. (P.) elatmae* that are similar in morphology (Nikitin, 1878, p. 56, Plate II, nos. 16, 17), the shells of *C. (P.) pishmae* are larger and show relatively less thick and higher whorls, the last of which has a high, steep, and smooth umbilical wall. At the same time, both species are similar in their ornamentation and in the cross-section shapes of inner whorls.

Distribution: lower Callovian, *C. pishmae* Beds, Pizhma River, Pechora River basin.

Cadoceras (Streptocadoceras) variabile Spath, 1932
Plate, nos. 2a and 2b

Cadoceras variabile (Spath, 1932, p. 75; Plate XVIII, no. 1; Plate XIX, nos. 1a–1c; Voronets, 1962. Plate 20, no. 3; Poulton, 1987, p. 56, Plate 27, no. 1; Meledina and Nalnyaeva, 1991, p. 74).

Cadoceras crassum (Frebold, 1961, p. 17; Plate XIV, no. 2; Plate XVII, no. 1).

Cadoceras (Streptocadoceras) variabile (Meledina, 1994; Plate VI, nos. 2a, 2b; Plate VIII, nos. 1a, 1b, 2a, 2b; Plate XII, no. 2).

Holotype (Spath, 1932, Plate XIX, nos. 1a–1c; British Museum; Vardecleft Formation, Bathonian ?, East Greenland).

Material: three nearly complete, though slightly deformed, casts.

Description. Shells are 85 to 95 mm in diameter and have slightly convex or flat lateral sides and clearly convex ventral sides. The whorl thickness is approximately equal to 1/2 of the shell diameter. Rounded-trapezoid cross sections of whorls are somewhat extended in width. The umbilicus is moderately wide (*U/D* = 30–33%). The umbilical wall is gently inclined. The dimension of umbilical hyponome that has a rounded inflection that measures up to 43–45% of the whorl diameter. The living chamber equals up to 4/5 of the whorl. The overhanging aperture ends with the contracted lip. Inner whorls are ornamented with bifid ribs, which have their origin at the upper part of umbilical wall. They are inclined up ahead and form elongated tubercles at the inflection. Among them, there are rare inset ribs. At the lateral sides, ribs blur toward the living chamber, though they are sometimes preserved on its walls as well. Number of tubercles is 19–20 over the last whorl and 21–24 over inner whorls.

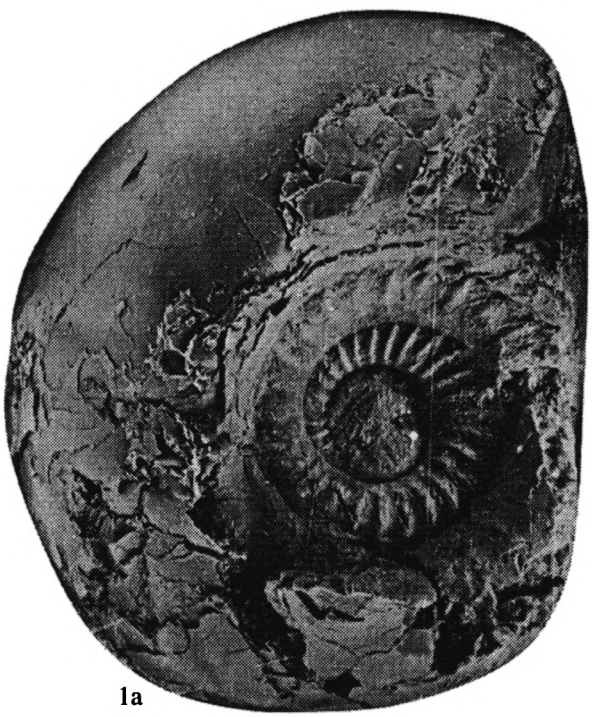
Comparison. As compared to *C. (S.) subtenuicostatum* Vor. (Voronets, 1962, p. 53; Plate XVII, nos. 1a, 1b, 2a, 2b, 3a, 3b; Plate XX, nos. 2a, 2b), *C. (S.) variabile* Spath is ornamented over phragmocone with thicker and less frequent ribs usually blurring at the living chamber. Cross sections of its whorls are less thick and lower (*T/D* = 47–50% versus 45–65% and *H/D* = 35–40% versus 37–45%), and umbilical inflection in the living chamber is more rounded.

In the form and size of umbilicus and umbilical hyponome, *C. (S.) variabile* resembles *C. (S.) bodylevskyi* (Frebold, 1961, p. 10; Plate XII, nos. 1a–1c; Plate XIX, nos. 1, 2). However, it differs from the latter in having the more flat (up to 7–20%) shell, lower and wider whorl sections, and better rounded umbilical inflection.

Distribution. Upper Bathonian: *C. variabile* Beds at the Pizhma River (Pechora River basin); upper part of

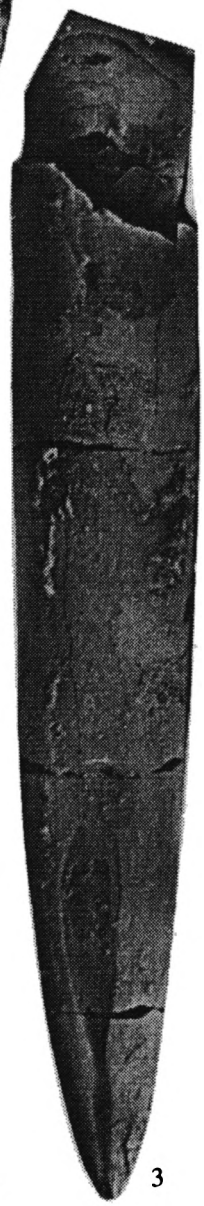
(1) *Cadoceras (Paracadoceras) pishmae* Meledina, specimen 489-263; (1a) lateral side, (1b) ventral side; Pizhma River, Exposure 14, Bed 1; lower Callovian, *C. pishmae* Beds. (2) *Cadoceras (Streptocadoceras) variabile* Spath, specimen 489-214; (2a) lateral side, (2b) ventral side; Pizhma River, Exposure 13, Bed. 1; upper Bathonian, *C. variabile* Beds. (3) *Pachyteuthis (Pachyteuthis) tschernyschewi* (Krimholz), specimen 89-1, ventral side; Izhma River, Dreshchanka Creek; middle Bathonian, *Arcticoceras ishmae–A. harlandi* Beds; described by Sachs and Nal'nyaeva (1966, p. 23). (4) *Paramegateuthis ishmensis* (Gustomesov), specimen 89-4, dorsal side; the same locality and age; described by Sachs and Nal'nyaeva (1975, p. 64). (5, 6) *Belemnopsis* sp. ind. (*B.* ? ex gr. *latisulcatus* Orb.); (5) specimen 89-2, (5a) ventral side, (5b) dorsal side; (6) specimen 89-3, (6a) ventral side, (6b) dorsal side; the same locality and age.

Plate



1a

1b



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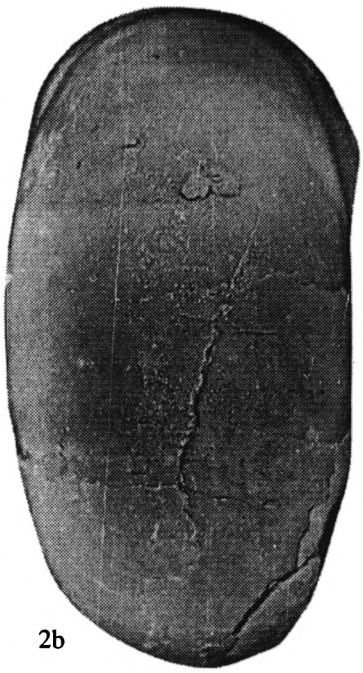
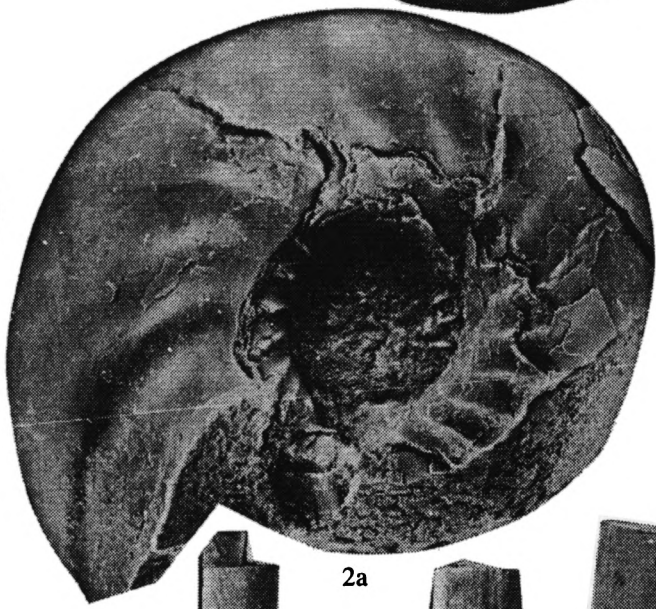


2b



2a

4

5a

5b

6a

6b

Measured parameters of the shell (mm and percentage relative to diameter in parentheses)

Specimen no.	Diameter, *D*	Height of whorl, *H*	Thickness of whorl, *T*	Diameter of umbilicus, *U*	Diameter of umbilical hyponome, *UC*	*T/H* ratio	Number of tubercles
489–214	93	33(35.4)	46(49.4)	28(30)	42(45)	1.4	19
489–506	87	–	–	–	–	–	–
	83	33(40)	39(47)	27(32.5)	38(45.7)	1.2	20
	73	27(37)	38(52)	22(30)	32(44)	1.4	24
489–508	85	34(40)	40(47)	28(33)	39(43)	1.1	20

Cadoceras barnstoni Zone in East Siberia (*C. variabile* Beds); *C. variabile* Zone in East Greenland; and *C. barnstoni* Zone in western Canada.

Description of Belemnites

Belemnopsis sp. ind. (*Belemnopsis* ? ex gr. *latesulcatus* Orb.)
Plate nos. 5a, 5b, 6a, and 6b

Specimens represent the incomplete fusiform rostrums of a middle size. Their ventral side is decorated with a wide and shallow furrow extending over the entire rostrum. Cross sections are rounded and compressed in dorsal–ventral direction.

Rostrums, especially specimen 89-2, are comparable with those of *Belemnopsis latesulcatus* Orb. (Pugaczewska, 1961, p. 150, Plate XI, no. 12).

The distribution range of *Belemnopsis latesulcatus* corresponds to Bathonian–Callovian deposits of Germany, France, Switzerland, England, Poland, and the Caucasus. *Belemnopsis* sp. ind. (*B.* ? ex gr. *latesulcatus*) is detected in the middle Bathonian *Arcticoceras ishmae–A. harlandi* Beds at the Dreshchanka Creek locality of the Izhma River basin.

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New Data on Composition and Age of Malomikhailovka Flora from the Upper Cretaceous of the Amur River Lower Courses

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Abstract—The described flora is studied near the village of Malomikhailovka, where it is confined to synonymous formation of an intermontane depression at the northern end of the Sikhote Alin volcanic belt. In its dominant taxa and in the composition of plant groupings, the studied flora notably differs from the coeval Tsagayan flora of the Zeya–Bureya depression and from floras occurring in western Sakhalin and eastern Sikhote Alin. It includes a new genus, *Amurocyparis*, with type species *A. sokoloviae*, as well as *Larix puerensis* sp. nov., *Cryptomerites tutshkovii* sp. nov., *Corylites amurensis* sp. nov., and *Palaeocarpinus pacifica* sp. nov., which are described here. The latter taxon was discovered for the first time in Russia.

Key words: *flora, Danian, Amur River, tuffaceous–sedimentary deposits, ferns, horsetails, conifers, pollen assemblage, Amurocyparis.*

The Malomikhailovka flora was found in a tuffaceous-sedimentary rock member about 25 m thick, which is exposed at the base of a 6- to 8-m-high terrace on the right-hand side of the Amur River. This locality is situated between the abandoned villages of Malomikhailovka and Pat, where the northeastern spur of the Chayatyn Ridge approaches to the river on the other side of its valley (Fig. 1). Sokolova (1940) was the first to collect fossil plants here in 1934. Later, the geological structure of this locality and its flora were examined by I.I. Tuchkov, Z.P. Potapova, V.I. Sukhov, L.P. Botyleva, T.N. Baikovskaya, M.M. Koshman, V.A. Vakhrameev, M.A. Akhmet'ev, and G.M. Brattseva (Vakhrameev, 1960; Tuchkov, 1960; Vereshchagin and Potapova, 1966; Sukhov, 1967; Akhmet'ev *et al.*, 1976; Akhmet'ev, 1993). However, the taxonomic composition of this flora is still insufficiently studied, and the stratigraphic position of the host beds remains disputable. In contrast to other transitional Cretaceous–Paleogene floras buried either in perioceanic coastal lowlands, or in spacious intercontinental depressions of the Amur–Zeya type, this one is deposited in a small intermontane trough at the western periphery of the Sikhote Alin volcanic belt far from the ocean. It exemplifies an unusual type of vegetation, because the trough bottom has been located higher than in other areas. Moreover, the flora in question is perfectly preserved, unlike the other floras, and this situation proved favorable for detecting a series of new plants characteristic of the region in the past.

The detailed study of flora-bearing deposits and bed-by-bed sampling of plant remains were performed by Akhmet'ev *et al.* (1976). To the south of Malomikhailovka, the flora-bearing beds are exposed over a distance of several kilometers. They form a gentle synclinal fold striking northeastward and having dip angles of 10°–12° at its limbs (Fig. 2). The sequence consists of alternating volcanosedimentary and sedimentary rocks. It discordantly overlies the basement composed of variable extrusive and pyroclastic rocks represented by andesites in the north and by rhyolite to dacite varieties in the south. The fine-grained rocks of the tuffaceous–

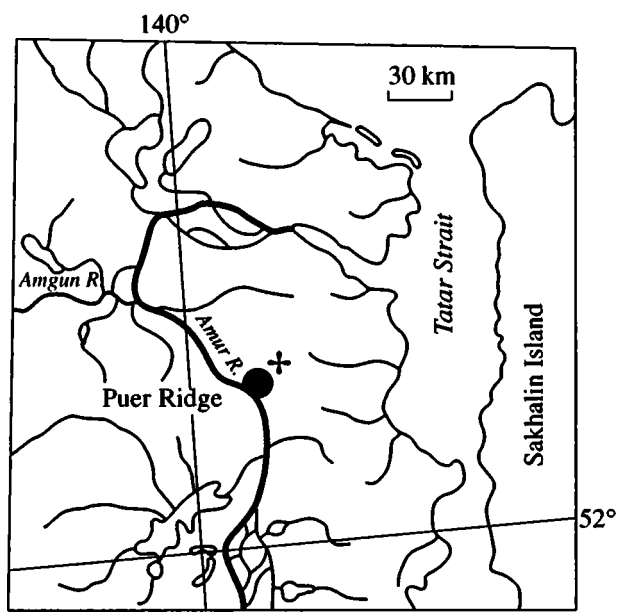


Fig. 1. Geographic locality of the Malomikhailovka flora.

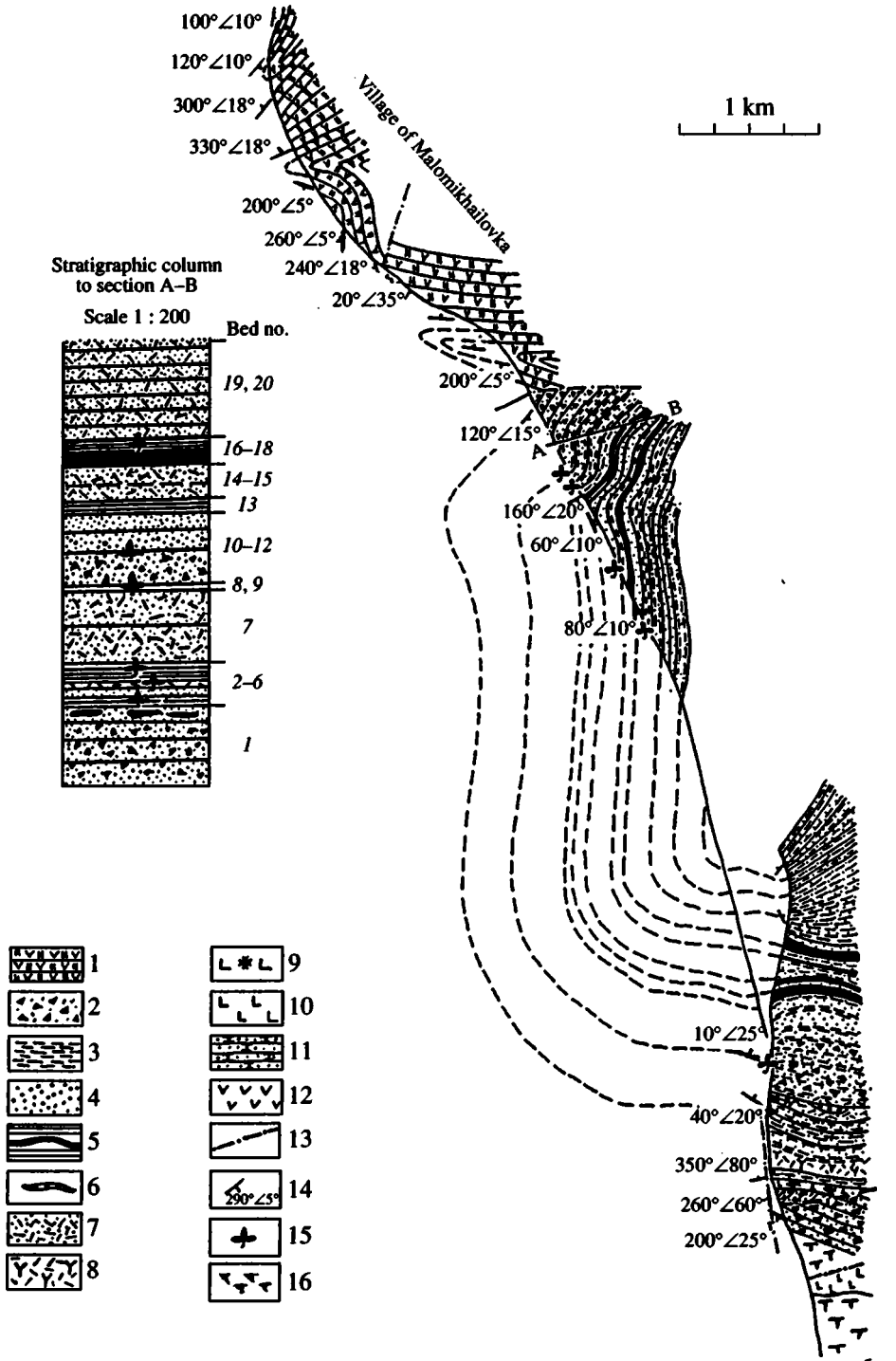


Fig. 2. Geological scheme of the Amur River bank, the Malomikhailovka–Pat area (compiled by M.A. Akhmet'ev and Yu.A. Ivanov): (1) andesite–dacite; (2) tuff and tuffaceous sandstone; (3) tuffite and tuffaceous siltstone; (4) fine- to medium-grained sandstone; (5) lignite and carbonaceous claystone; (6) lenses of tuffaceous conglomerate and gravel; (7) acidic tuff; (8) dacite; (9) amygdaloidal basaltic andesite; (10) basalt; (11) sandstone; (12) andesite; (13) fault; (14) bed attitude; (15) sampling points of fossil flora; (16) unconsolidated Quaternary deposits.

sedimentary member grade to the southeast into the coarse-grained beds of rhyolite tuffs. Andesites are estimated to be 74 Ma old (radiometric determination by T.G. Koval'chuk from the Far East State University, oral presentation in 1969). Tuchkov (1960), who distinguished the Malomikhailovka Formation, considered the tuffaceous-sedimentary member as its upper unit overlying the lower part, composed of andesites.

Akhmet'ev *et al.* (1976), who established the tectonic nature of contact between this member and andesites and traced the lateral transition of the former into beds of rhyolite composition, argued against the idea of uniting the first two rock types into a single formation.

The flora-bearing sequence shown in Fig. 2 consists of the following beds:

(1) tuffaceous sandstone, coarse-grained, indistinctly laminated, enclosing in the middle part an interlayer with remains of fossil wood (4 m);

(2) tuffite, light brown, grading upward into cherty tuffaceous siltstone of dark gray color (0.8 m);

(3) tuffaceous siltstone, dark gray, carbonaceous (0.15 m);

(4) tuffite, dark gray, intercalated with interlayers of tuffaceous sandstones locally as thick as 0.5–0.6 m (2–2.5 m);

(5) tuffaceous siltstone, dark gray, carbonaceous (0.2 m);

(6) tuffite, dark gray (0.5 m);

(7) tuffaceous sandstone, rusty brown, medium- to coarse-grained, alternating with tuffite and tuffaceous siltstone interlayers (3.2 m);

(8) tuffaceous siltstone, dark gray, carbonaceous (0.05 m);

(9) tuffite, dark gray (0.6 m);

(10) tuffaceous sandstone, brown, coarse-grained (1.3 m);

(11) tuffaceous claystone, dark gray, carbonaceous (0.12 m);

(12) tuffaceous sandstone, coarse-grained, enclosing interlayers of tuffaceous siltstone (1.5 m);

(13) tuffaceous claystone, dark gray, carbonaceous (0.15 m);

(14) tuffaceous sandstone, coarse-grained, with spheroid and loaflike jointing (2.5 m);

(15) tuff, brownish gray, fine-grained, intercalated with tuffite interlayers (1 m);

(16) tuffaceous siltstone, dark gray, carbonaceous (0.2 m);

(17) tuff, fine-grained (0.5 m);

(18) lignite (0.5 m);

(19) tuff, light gray, fine-grained (1 m);

(20) tuffaceous sandstone, dark brown, fine-grained, enclosing thin interlayers of carbonaceous claystone (3 m).

Lignite interlayers in carbonaceous siltstone and sandstone, together with structural and textural features of rocks and the types of fossil plants characteristic of boggy landscapes, indicate that flora-bearing beds have accumulated in an intermontane basin surrounded by volcanic edifices supplying the basin with volcanic material. Hydrological environments in the basin were variable. Lignite and carbonaceous siltstone beds mark periods of swamping, whereas lenticular interlayers of cross-bedded gravelstone and shingle point to alluvial environments of sedimentation.

Megafossils were encountered in beds 2, 6, 9, 11, 13, and 20. They were most abundant in Bed 2. The basal part of the bed yielded plentiful (more than 100 specimens) remains of horsetail species. Less abundant (dozens of specimens) are *Onoclea* fern fragments and Pinaceae megastrobiles. Interlayers of tuffaceous siltstone in the bed contain hundreds of *Metasequoia* shoot remains and are only somewhat less rich in specimens of *Fokieniopsis, Cryptomerites, Amurocyparis,* and *Dennstaedtia.* These taxa coexist with considerably less abundant remains of *Onoclea,* mosses, micro- and megastrobiles of Taxodiaceae, and *Corylites* leaves.

Bed 6 is also saturated with fossil plants. *Metasequoia* shoots (hundreds of specimens) and *Corylites* leaves (dozens of specimens) are dominant at this level. Less abundant (10–15 specimens) are ferns *Onoclea* and *Dennstaedtia, Amurocyparis* shoots, and remains of horsetails. We also detected two leaves of *Trochodendroides* ex gr. *arctica* (Heer) Berry.

Bed 9 yielded only 10–15 specimens of poorly preserved remains of *Corylites* and *Metasequoia.*

Bed 11, especially its basal portion, turned out to be rich in fossil flora. Above the contact with underlying coarse-grained sandstone, we found here remains of horsetails, several impressions of *Ginkgo* leaves, more than ten impressions of *Nyssidium* fruits, and abundant *Trochodendroides* leaves and *Metasequoia* shoots.

Bed 13 contains plentiful remains of *Onoclea,* and the tuffaceous sandstone of Bed 20 bears fragmented remains of horsetails.

The assemblage of *Corylites* and *Metasequoia* with accessory *Amurocyparis, Fokieniopsis,* and *Dennstaedtia,* which is confined to carbonaceous tuffite beds, is most persistent. Accordingly, it may correspond to the most characteristic type of past vegetation around a small water basin. Fossil horsetails and *Onoclea* ferns mainly form the monodominant burials confined to coarse-grained rocks, in the first case, and to fine-grained sediments, in the second. In general, this is concordant with present-day habitat environments of these plants, as some horsetail species often populate the recently exsiccated shoals, and *Onoclea* overgrowths are typical of marshy lowlands. The discrete occurrence of *Corylites* and *Trochodendroides* leaves is particularly interesting. Both taxa were clearly dominant and coexisted with *Metasequoia,* although the abundant *Trochodendroides* remains are concentrated together with *Ginkgo* leaves in the coarse-grained rocks at the point of contact between beds 10 and 11. It is quite possible that these latter two species populated some elevated areas of embankment type with well-aerated soils. In other coeval flora localities of the same lithologic facies, no such an interrelation has been observed. Remains of *Cryptomerites, Larix* strobiles, and coniferous needles of other indeterminable Pinaceae are less abundant. These, especially *Larix*

specimens, are presumable elements of upland and slope vegetation formerly located at some distance from the water basin in question.

In total, the Malomikhailovka flora includes the following taxa: *Equisetum* sp., *Muscites* sp., *Ginkgo* ex gr. *adiantoides* (Unger) Heer, *Asplenium* sp., *Onoclea hebridica* (Forbes) Johnson, *Dennstaedtia tschuktschorum* Krysht., *Metasequoia occidentalis* (Newb.) Chaney, *Fokieniopsis catenulata* (Bell) McIver et Basinger, *Cryptomerites tutshkovii* sp. nov., *Amurocyparis sokoloviae* gen. et sp. nov., *Pityospermum* sp., *Pityostrobus* sp., *Larix puerensis* sp. nov., *Corylites amurensis* sp. nov., *Palaeocarpinus pacifica* sp. nov., *Trochodendroides* ex gr. *arctica* (Heer) Berry, *Trochodendroides* sp., and *Nyssidium arcticum* (Krysht.) Iljinsk. Plants dominant flora of limited diversity include *Corylites, Metasequoia, Trochodendroides, Nyssidium, Onoclea, Dennstaedtia, Fokieniopsis,* and *Ginkgo* genera, all very characteristic of floral assemblages of the terminal Cretaceous–initial Paleogene period. Many of them display wide geographic ranges and a high degree of individual variability. Apart from *Trochodendroides* ex gr. *arctica*, the Malomikhailovka Formation yielded several other leaf specimens of this genus, which are elongated and rimmed by frequent sharp dens (Fig. 3, no. 8). Impressions of *Corylites amurensis* leaves occur together with fruits of *Palaeocarpinus pacifica,* discovered for the first time in Russia. Shoots of *Amurocyparis* and *Larix* bearing female strobiles are of particular scientific significance. Remains of *Larix puerensis* represent the oldest known species of this genus at present. The discovered generative organs of Cupressaceae are extremely rare in Cretaceous–Paleogene deposits and highly important for understanding the evolution of this family.

In its composition, the Malomikhailovka flora most closely resembles the Danian Boshnyakovo flora of Sakhalin (Krasilov, 1979), the late Maastrichtian–Danian Rarytkin flora of Koryak Upland (Golovneva, 1994), and the Maastrichtian Samarga flora of Sikhote Alin (Oleinikov and Nevolina, 1985). All these floras display a predominance of *Metasequoia* and different *Corylus* species, in association with *Trochodendroides* remains. The second taxon is represented by *Corylus beringiana* (Krysht.) Golovn. in the Rarytkin flora, by *Corylus (Corylites) protoinsignis* Krasilov in the Boshnyakovo assemblage, and by *Corylus* sp. in the Samarga flora. The latter, which represents former vegetation of the Sikhote Alin volcanic belt, is most similar to the Malomikhailovka flora. Characteristic conifers of the Samarga flora are *Metasequoia* and typical Cretaceous taxa *Sequoia* sp. and *Thuja cretacea* Newb. At the same time, this flora is rich in a different taxon of Betulaceae and includes some old taxa of angiosperms. The Samarga flora, similar in composition of megafossils to Danian floras, is attributed to the Maastrichtian on the basis of palynological spectra and radiological dates. According to the latter, the Samarga Horizon could be as old as 50 to 75 Ma (Oleinikov, 1978; Filimonova, 1983; Mikhailov and Gracheva, 1987; Matyunin, 1988). This interval corresponds to both the late Campanian–Maastrichtian and the Danian, if the Cretaceous–Paleogene boundary is 65 Ma old (Harland *et al.,* 1990). However, pollen assemblages from the Samarga Formation show a preponderance of *Aquilapollenites, Mancicorpus, Wodehouseia, Orbiculapollis,* and *Ulmoidepites* genera, which is typical of the Maastrichtian (Markevich, 1987). In addition to the Malomikhailovka flora from the Amur River lower courses, the *Corylites amurensis–Metasequoia* plant assemblage is detected in the upper part of the Levaya Sobolevka Formation, occupying a higher stratigraphic level as compared to the Samarga Formation. Palynospectra from the Levaya Sobolevka Formation are considerably depleted in Maastrichtian taxa and display a predominance of pollen of Betulaceae and a Miricaceae affinity. Diverse pollen of flowering plants is replaced here by dominant Taxodiaceae in association with rather abundant fern spores. These pollen spectra are dated back to the early Danian or late Maastrichtian–early Danian. Pollen assemblages from flora-bearing beds of the Malomikhailovka Formation, which are similar to the above spectra in composition, are also the early Paleocene in age (Akhmet'ev *et al.,* 1976).

Thus, the data suggest that the age interval of the Malomikhailovka flora spans the Danian and probably the uppermost Maastrichtian. In palynological aspect, this flora is most similar to the upper Tsagayan, Boshnyakovo, and upper Rarytkin floras. It should be mentioned that the Danian age has been suggested for the Malomikhailovka flora from the very beginning (Sokolova, 1940), when it was correlated with the Tsagayan flora of the Amur region, though the latter represents a somewhat different type of Danian floras and demonstrates a predominance of the *Trochodendroides–Taxodium* assemblage, instead of the *Corylites–Metasequoia* group. At present, the indicated age interval of the Malomikhailovka flora is substantiated by results of stratigraphic and paleobotanical data obtained by various researchers.

A description of several new taxa distinguished in the Malomikhailovka flora is presented below. They all are from the fossil plant collection no. 3808, assembled by V.A. Vakhrameev and M.A. Akhmet'ev in 1973 and stored at the Laboratory of Paleobotany of the Geological Institute, Russian Academy of Sciences.

Family Cupressaceae

Genus *Amurocyparis* Golovn. et Akhmetiev, gen. nov.

Genus name is after the Amur River.

Type species: Amurocyparis sokoloviae sp. nov.; Danian, Malomikhailovka Formation, Amur River (lower courses).

Diagnosis. Shoots are branching in a single plane, mainly in an alternate, and sometimes in a decussate, arrangement. Leaves are lepidoid. Facial leaves are flat

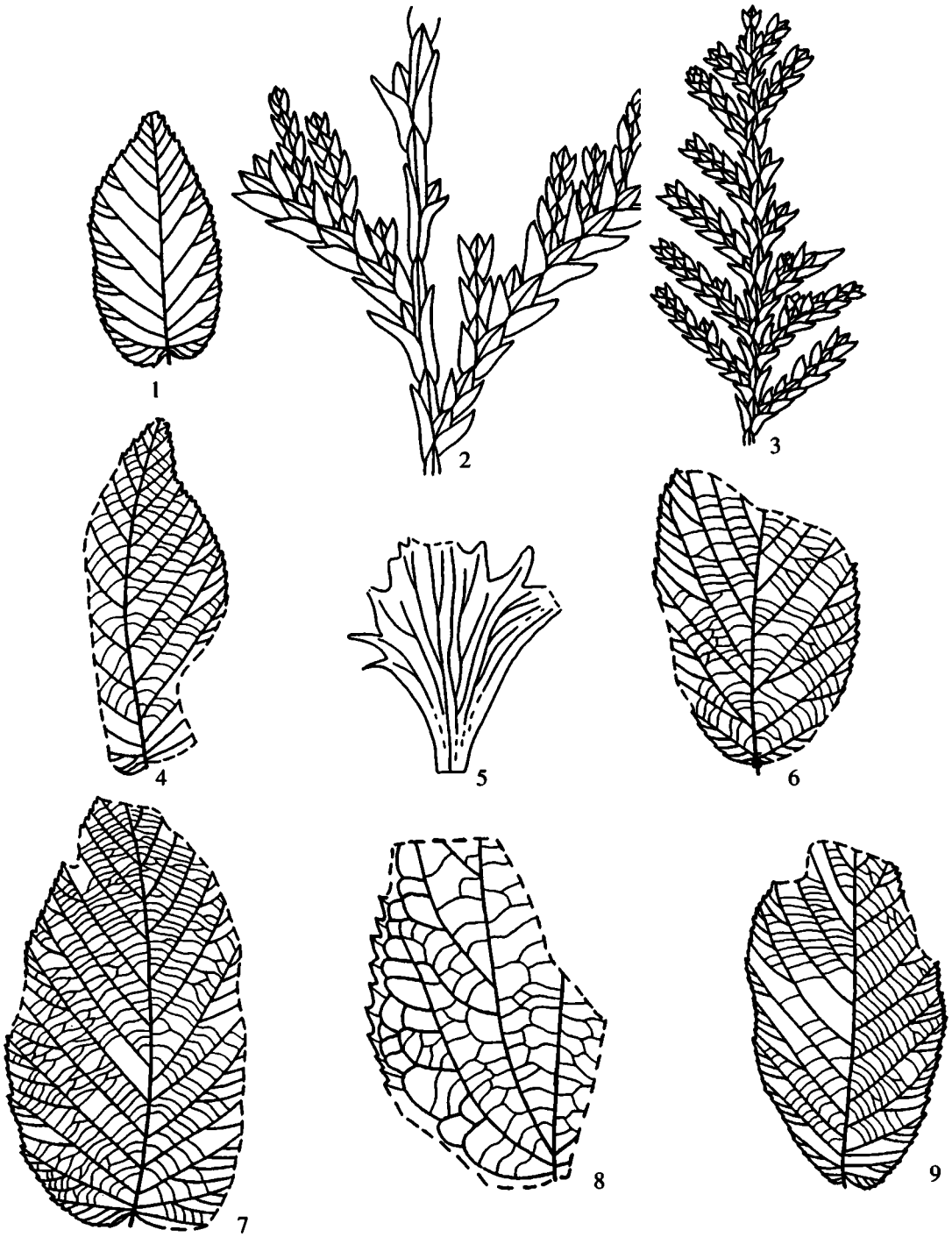


Fig. 3. Fossil Danian plants from the Malomikhailovka Formation and the Amur River lower courses: (1) leaf of *Corylites amurensis* Golovn. et Akhmetiev, specimen no. 3808/121, ×1; (2, 3) *Amurocyparis sokoloviae*, paratype, specimen no. 110, ×2 (2), and specimen no. 389, ×2 (3); (4) leaf of *Corylites amurensis* Golovn. et Akhmetiev, specimen no. 3808/603, ×1; (5) *Palaeocarpinus pacifica*, holotype, specimen no. 3c, ×7; (6, 7) leaves of *Corylites amurensis* Golovn. et Akhmetiev, specimen no. 3808/116, ×1 (6) and specimen no. 3808/295a, ×1 (7); (8) *Trochodendroides* sp., specimen no. 3808/ 394, ×1; (9) leaf of *Corylites amurensis* Golovn. et Akhmetiev, specimen no. 3808/370, ×1.

and pressed against the stem; their upper ends overlap the bases of higher lateral leaves, which are falculate in shape and have a wide base and acuminate free apex.

Semeniferous strobiles are rounded and small (2.5–3 mm in diameter); they occupy terminal positions at short foliaceous lateral shoots set alternately or at one side of the branch in the base of lateral leaves. After dissemination, strobiles remain to be fixed at the branch. They consist of four ligneous squamae of obtriangular or rounded-rhomboid form. Microstrobiles of about

1 mm in diameter consist of several squama pairs and rest at the ends of short acroscopic sprays.

Species composition: A. sokoloviae (type species); *A. corpulenta* (Bell) Golovn. [=*Thuites corpulentus* Bell, 1957, p. 35, Plate 18, no. 8; =*Chamaecyparis corpulenta* (Bell) McIver, 1994]; Santonian, Comox Formation, Vancouver Island.

Comparison. In morphology of leaves, *Amurocyparis* shoots are most similar to those of Cupressaceae tribes Thujopsideae and Cupresseae. The leafing of *A. corpulenta* branches resembles that typical of the genus *Thuja.* In both cases, lateral leaves are falculate and pressed against shoots, leaving only the acuminate apex free. Lateral leaves of *A. sokoloviae* have long free ends. Similar shoots are characteristic of genus *Chamaecyparis,* for instance, *C. nootkatensis* (D. Don) Spach.; however, the decussate ramification is nontypical of the tribes Thujopsideae and Cupresseae. This type of branching is characteristic of representatives of the Libocedreae tribe from the southern hemisphere and of the fossil genus *Mesocyparis* from the Maastrichtian and Paleocene deposits of northeastern Russia, Canada, and Spitsbergen (McIver and Basinger, 1987; Golovneva, 1988, 1994; McIver and Aulenback, 1994). Some representatives of Libocedreae and Maastrichtian *Mesocyparis umbonata* McIver et Aulenback display a combination of decussate and alternate ramification (McIver, 1994). Fragments of decussate shoots of *A. sokoloviae* are almost identical to those of *Mesocyparis.* At the same time, if decussate ramification is the rule for *M. umbonata* and the alternate shoot arrangement is the exception, the last type of branching is, in contrast, typical of *Amurocyparis* species, and the number of their shoots with decussate ramification is below 5%.

Seminiferous strobiles of *Amurocyparis* closely resemble those of *Mesocyparis.* They are equally small and consist of four ligneous squamae, each presumably bearing two seeds. However, assidenous strobiles of *Mesocyparis* are arranged as series of decussate pairs at special aphyllous twigs, whereas the terminal strobiles of *Amurocyparis* are situated at short lateral leaf-bearing twigs branching either in the alternate manner, or from one side of a shoot. Strobile arrangements of this kind are typical of many Cupressaceae. Among present-day species of this family, only *Microbiota decussata* Komarov, from the Thujopsideae tribe and growing in the Sikhote Alin Ridge, has small four-squama strobiles; however, their squamae are coriaceous but not ligneous, and each strobile bears only one seed.

McIver (1994) believes that strobiles of the *Amurocyparis* type are characteristic of the genus *Chamaecyparis,* and he refers their fossil remains to this taxon. This is not entirely correct, because strobiles of *Chamaecyparis* are notably larger and covered with numerous peltate squamae.

In our opinion, the structure and arrangement of *Amurocyparis* microstrobiles are almost identical to these morphological elements of *Mesocyparis.*

Thus, the fossil genus *Amurocyparis* is most similar to *Mesocyparis* in the ramification style and structure of seminiferous strobiles. In morphology of shoots and leaves, as well as in strobile arrangement, it shows an affinity with the taxa of Thujopsideae and Cupresseae tribes. Morphological features of *Mesocyparis* can be found in both Thujopsideae and Libocedreae forms. In contrast to this genus, *Amurocyparis* is much closer to Thujopsideae and other Cupressaceae from the northern hemisphere.

Amurocyparis sokoloviae Golovn. et Akhmetiev, sp. nov.
Plate I, nos. 1–4, 7–10, and 12; Plate II, no. 1; Fig. 3, nos. 2 and 3

Androvettia catenulata Bell (Akhmet'ev, 1993; Plate III, nos. 4, 5, 7, and 13; Plate IX, no. 1)

Species name given in honor of geologist E.I. Sokolova, who discovered the Malomikhailovka flora locality.

Holotype: specimen no. 3808/600 in collection of Geological Institute, Russian Academy of Sciences; seminiferous strobiles, Malomikhailovka Formation, Danian, Plate I, no. 2.

Paratype: specimen no. 3808/110 in collection of Geological Institute, Russian Academy of Sciences; leaf-bearing shoots, Malomikhailovka Formation, Danian; Plate I, no. 10; Fig. 3, no. 2.

Description. Our collection includes small shoots with ramifications of three orders and less. Ramification, which is predominantly alternate and rarely decussate, takes place in a single plane. Offshoots of penultimate order in larger shoots are often asymmetrically branching: 5–6 terminal twigs are situated at the acroscopic side, whereas basiscopic side has either 1–2 or no terminal twigs. Leaf arrangement is decussate. Facial leaves are long, narrow, and pressed against the stem; their carina is weakly pronounced, and the acute apex either gradually narrowing, or rhomboid in shape, overlapping the base of the higher lateral leaf. The lateral leaves have a wide base and narrowing acute apex; they are ether straight or slightly falculate in form. Free parts of lateral leaves deviate from the stem approximately for a half of their length. Terminal twigs 0.6–1.5 cm long consist of 3–7 pairs of lateral leaves. Penultimate offshoots are 1.5–5 cm long and unite 7 to 20 pairs of lateral leaves. Lateral and facial leaves of large axial shoots with extended internodes are more elongated and narrow; they can be 2–3 times longer than leaves of the terminal and penultimate offshoots. The leaf length varies from 1.5–2.5 mm at the terminal twigs to 3–4 and 5–8 mm at the penultimate and long axial shoots respectively.

About 5% of all shoots display decussate ramification. In this case, the final lateral offshoots have their origin in each base of lateral leaves and usually consist of 2–4 pairs of new lateral leaves. One of the specimens has rounded microstrobiles (0.8–0.9 mm in diameter), which rest in the base of lateral leaves at the acroscopic

side of decussate lateral offshoots and consist of several squama pairs. Large, normally evolved shoots with decussate ramification are absent.

Rounded and small female strobiles (2.5–3 mm in diameter) are situated at the ends of short lateral twigs bearing 3 to 4 pairs of dimorphic decussate leaves either alternating or located at one side of the twig. Squamae are oval-obtriangular or oval-rhomboid in shape; they have the brachyacute apex sometimes ornamented with small dens. There are two seed impressions in inner sides of squamae.

Comparison. New species differs from *Amurocyparis corpulenta* (Bell) comb. nov. chiefly in morphology of leaves. Lateral leaves of *A. corpulenta* are short and tightly envelope facial leaves mainly without free ends. Lateral leaves of *A. sokoloviae* are longer and diverge out of shoots approximately beginning from their middle part.

Distribution: Malomikhailovka Formation, Amur River lower courses, Danian.

Material: seminiferous strobiles (specimens 3a, 62, 257, 308a, 387, and 600); microstrobiles (specimen 257); vegetative shoots (specimens 30, 47, 64, 71, 77, 110, 131, 143c, 246, 309, 312, 369, 371, 382a, 389, 390, 400, and others).

Family Taxodiaceae (?)

Cryptomerites tutshkovii Golovn. et Akhmetiev, sp. nov.
Plate II, no. 4

Sequoia sp. (Akhmet'ev, 1993, Plate IV, nos. 11, 13, and 14)

Species name is given in honor of geologist I.I. Tuchkov.

Holotype: specimen no. 3808/519 in collection of Geological Institute, Russian Academy of Sciences; Malomikhailovka Formation, Danian; Plate II, no. 4.

Description. Ramification of shoots is alternate, and arrangement of leaves is spiral. In their cross sections, leaves are approximately tetrahedral, laterally flattened. They are straight or slightly falculate in shape and acute at the apex; the leaf base is decurrent toward the shoot axis. Leaf is 3–8 mm long and about 1.5 mm wide.

Comparison. Shoots of new species differ from those of *C. ancistrophyllus* Golovn. from Maastrichtian deposits of the Koryak Upland (Golovneva, 1994) in having wider and longer leaves, which are almost straight, but not hamiform, and sparsely arranged at the shoot. The present-day genus *Cryptomeria* has similar tetrahedral decurrent leaves, but, in contrast to ancient species, its leaves are longer and thinner. In addition, they are more densely arranged at the shoot. The oldest forms of *Cryptomeria,* whose affinity to this genus is established on the basis of epidermis, are known to occur since the Cenomanian in northeastern Asia (Samylina, 1988).

Distribution: Malomikhailovka Formation, Amur River lower courses, Danian.

Material: specimens 519 and 215.

Family Pinaceae

Larix puerensis Golovn. et Akhmetiev, sp. nov.
Plate I, no. 11; Plate II, nos. 2 and 6

Species name is after the Puer Ridge in the Amur River lower courses.

Holotype: specimen no. 3808/601 in collection of Geological Institute, Russian Academy of Sciences; long shoot with short offshoots and strobile; Plate I, no. 11.

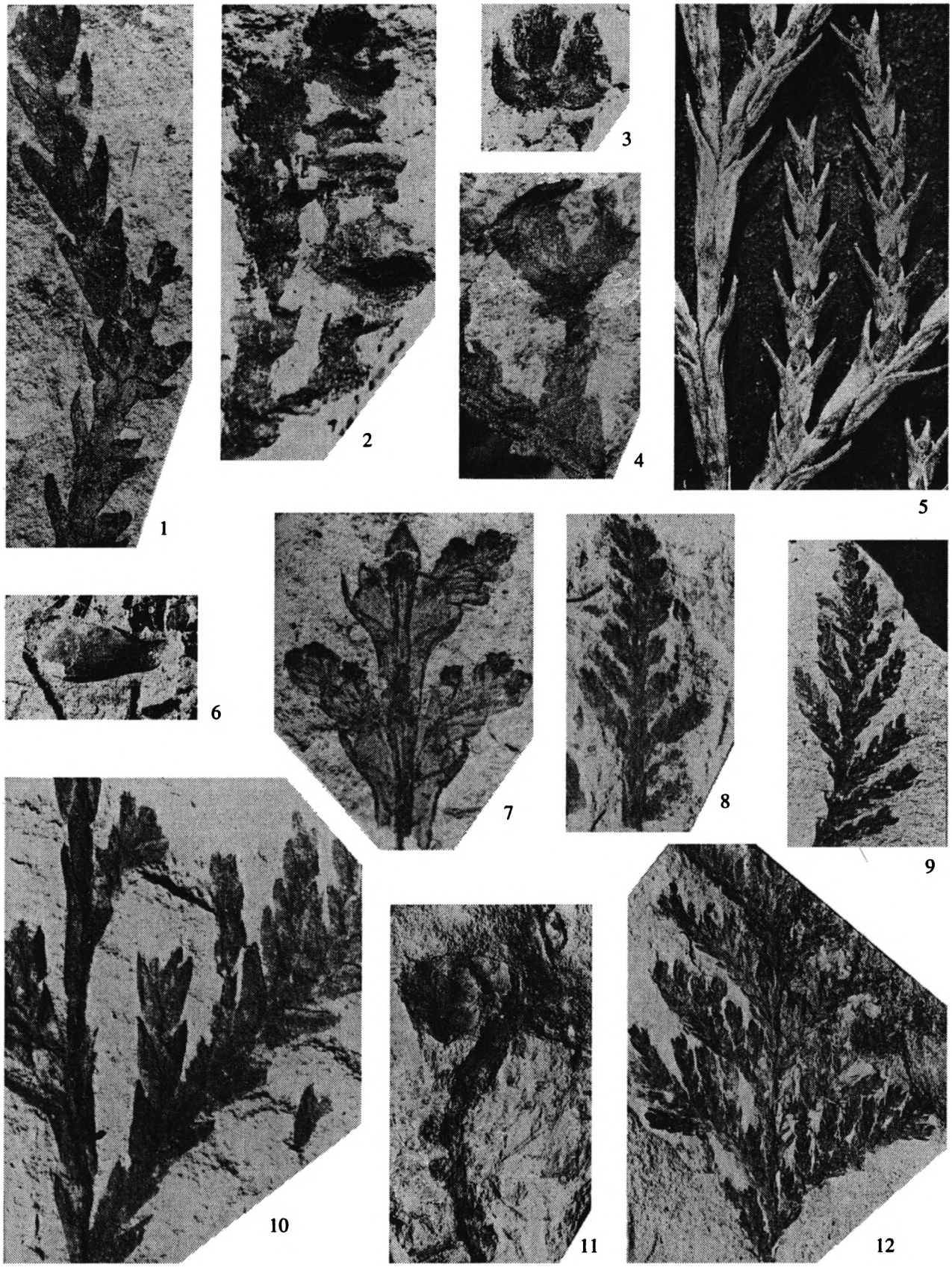
Paratype: specimen no. 3808/36 in collection of Geological Institute, Russian Academy of Sciences; leaves; Plate II, no. 6.

Description. Specimen represents a long shoot 3–4 mm in diameter with insertion points of solitary leaves and with short offshoots arranged in a spiral manner. The latter are elongated-oval or egg-shaped, 3–6 mm long, and 3–4 mm in diameter. In their apical parts, they are covered with small, tightly assembled scales. The egg-shaped or elliptical female strobiles are up to 2.5 cm long and 1.5 cm in diameter; they bear numerous coriaceous squamae and crown short offshoots. Seminiferous squamae are about 1 cm long, rounded at the base, and obtuse at the apex.

Deciduous leaves are flat and linear, up to 6 cm long, and 1–1.5 mm wide. They are gradually narrowing toward the base and rounded to brachyacute at the apex.

Comparison and remarks. Species from the Malomikhailovka Formation is the oldest representative of genus *Larix,* described on the basis of leaves and strobiles. *Larix* species usually occur in the temperate northern latitudes beginning from the Oligocene, and remains of *Larix altoborealis* Le Page et Basinger from the middle–upper Eocene deposits of Arctic Canada used to be considered as the most ancient specimens of the genus (Le Page and Basinger, 1991). The species in question differs from *L. altoborealis* in having longer and wider leaves, and also in the shape of seminiferous squamae. Squamae of *L. puerensis* are wider and truncated at the top, whereas these elements of *L. altoborealis* are oval, or almost circular in shape. Unfortunately, coating squamae, which are important in taxonomy of *Larix* genus, have not been preserved in impressions of *L. puerensis* strobiles. It is well known that the morphological similarity of leaves and strobiles is characteristic of *Larix* species that differ substantially in other aspects. Accordingly, our specimen is inappropriate for a detailed comparison with younger taxa. Despite its ancient features, the material we have is represented only by impressions, thus being inappropriate for elucidating the evolutionary history of *Larix* genus in contrast to *L. altoborealis*, whose remains include the mummified shoots, strobiles, and wood.

Plate I



Rather long leaves (5–6 cm in average) represent one of characteristic features of *L. puerensis*. Among the present-day larches, *L. olgensis* Henry, widespread in northeastern China, has similar leaves, acute at the apex and gradually narrowing toward the base. Some researchers (Ostenfeld and Larsen, 1930) consider this form only as a subspecies of *L. gmelini*. Leaves of this species are generally shorter (3–4 cm), but they also can be as long as 5–6 cm. The leaves we refer to *L. puerensis* have been found separately from shoots, however they are very similar to leaves of recent larches growing not far from the locality under consideration, and this is a good opportunity to suggest that the studied leaves and shoots with strobiles originated from one plant. In their size and form, these leaves resemble those of *Pseudolarix*, although in the latter case they are usually larger (up to 10–13 cm long and 5 mm wide). Moreover, we did not detect any other type of *Pseudolarix* remains in the Malomikhailovka Formation. In addition to strobiles and leaves from this formation, there are impressions of alate seeds of the *Larix* genus affinity (Plate II, no. 3). However, their precise identification is nearly impossible, because the Malomikhailovka flora includes several representatives of Pinaceae (Akhmet'ev, 1993).

Distribution: Malomikhailovka Formation, Amur River lower courses, Danian.

Material: seminiferous strobiles (specimens 376, 463, and 601); vegetative shoots (specimens 36, 312, and 3826).

Family Betulaceae

Corylites amurensis Golovn. et Akhmetiev, sp. nov.
Plate II, nos. 5 and 8–10; Fig. 3, nos. 1, 4, 6, 7, and 9

Alnites sp. (Akhmet'ev, 1993; Plate VII, nos. 4 and 7; Plate VIII, nos. 1 and 4; Plate IX, nos. 8 and 9).

Holotype: specimen no. 3808/385 in collection of Geological Institute, Russian Academy of Sciences; Malomikhailovka Formation, Danian; Plate II, no. 10.

Description. Leaves are 5–10 cm long and 2.5–5.5 cm wide, elliptical to wide oval- or egg-shaped. They display various degrees of asymmetry and have an extended acute apex and rounded to slightly cordate base. The leaf edge is serrate. Serratures are small (1–2 mm high), frequent, and triangular to galeate in shape; their apex is somewhat extended. Venation is of the pinnate-craspedodromous type. Secondary veins consisting of 14 to 19 pairs are straight to slightly curved. Three to four pairs of secondary veins usually are tightly arranged and almost perpendicular to the longitudinal vein. Upper secondary veins are more widely spaced and depart from the longitudinal vein with angles less than 30°–59°. In the middle part of a leaf, they are most long and have 4 to 6 basiscopic offsets. Venation of the third order is represented by frequent and curved anastomosing scalas. Venation of higher orders is well developed in a form of reticulum.

Comparison and remarks. Genus *Corylites* was first identified in the Paleocene flora of the Mall Island in Scotland (Gardner, 1887). In the course of subsequent, more comprehensive investigations of this flora (Boulter and Kvaček, 1989; Seward and Holttum, 1924), diagnostic features of the genus have been revised. According to the results obtained, characteristic features of *Corylites* leaves include numerous secondary veins (up to 17 pairs), even serratulate edges, and shortened narrow-spaced veins in the leaf base. Similar leaf peculiarities are characteristic of present-day *Alnus*, *Carpinus*, and some *Corylus* species. All generative organs of Betulaceae affinity useful for elucidating the taxonomic position of *Corylites* genus are missing in the Mall Island flora.

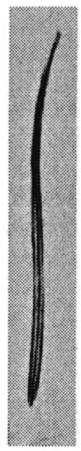
Because the morphology of Betulaceae leaves from the Malomikhailovka flora fits the diagnosis of *Corylites* species, they have been included into this taxon. In contrast to type species *C. hebridicus* Seward et Holttum, widespread in Paleocene deposits of Scotland and Spitsbergen, leaves of *C. amurensis* are generally lesser in size (Boulter and Kvaček, 1989); their secondary veins are more distinctly curved, and marginal serratures are smaller. Other representatives of Betulaceae from the Maastrichtian and Paleocene floras of northeastern Asia display far fewer lateral veins in leaves, and the arrangement of these veins differs from that of *Corylites* forms. Nevertheless, many such species have been attributed to this genus (Krasilov, 1979; Herman and Spicer, 1995; Manchester and Guo, 1996). Such a situation illustrates a tendency of including ancient Betulaceae variable in morphology into the *Corylites* genus, when their taxonomic position relative to recent taxa is unclear. This approach can be hardly accepted as correct, because it blurs distinction between genera *Corylus* and *Corylites*, and the latter, thus being morphologically boundless, turns out to be transformed into the composite taxon. The practice of transposing species from genus *Corylus* to genus *Corylites* by no

Plate I. Danian plants from the Malomikhailovka locality (the Amur River lower courses). (1–4) *Amurocyparis sokoloviae:* (1) shoot, specimen no. 369, ×5; (2) shoot with strobiles, holotype, specimen no. 600, ×6; (3) strobile, specimen no. 62, ×6; (4) shoot with strobile, specimen no. 3a, ×6. (5) *Chamaecyparis nootkatensis,* shoot, ×3, specimen from herbarium of the Botanical Institute, Russian Academy of Sciences. (6) *Pityospermum* sp., specimen no. 254, ×2.5. (7–10) *Amurocyparis sokoloviae:* (7) shoot with microstrobiles, specimen no. 257, ×4.5; (8) shoot with decussate ramification, specimen no. 604, ×3; (9) shoot with alternate ramification, specimen no. 371, ×2; (10) shoot, paratype, specimen no. 110, ×3. (11) *Larix puerensis,* shoot with strobile, holotype, specimen no. 601, ×1. (12) *Amurocyparis sokoloviae,* shoot, specimen no. 30, ×2.

Plate II

1

2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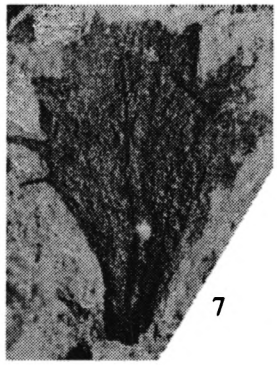
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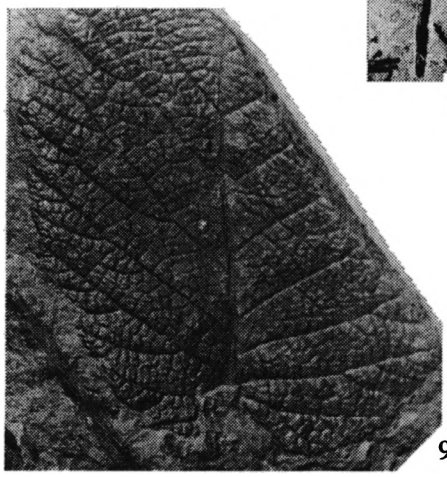
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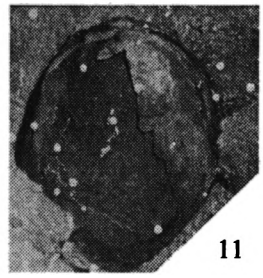
11

means helps to get insight into their taxonomic position, and that of creating the composite taxon is useless for understanding the diversity of ancient Betulaceae.

Leaves of *Corylites protoinsignis* Krassilov from the Danian Boshnyakovo flora of Sakhalin have the digitipinnate venation, lobulate margin, and 6 to 7 pairs of bow-shaped secondary veins (Krasilov, 1979).

According to Herman and Spicer (1995), *Corylites beringianus* (Krysht.) Herman [=*Corylus beringianum* (Krysht.) Golovn.) from the Maastrichtian–Danian deposits of the Rarytkin and Koryak formations of the Koryak Upland shows a high variability of leaf morphology (Golovneva, 1994). Despite the even serratulate margin and rather numerous secondary veins (10–15 pairs), many leaves of this species also display the digitipinnate type of venation and have well-developed basal veins with abundant offspring. Some morphotypes of this species, such as *carpinifolia* and *viburnifolia,* have leaves with pinnate venation and short basal veins, but these leaves have no narrow-spaced veins at their base.

Leaves of *Corylites bellatula* (Guo et Li) Manchester et Guo (=*Corylus bellatula* Guo et Li) from the Paleocene flora of northeastern China (Manchester and Guo, 1995) also show the pinnate venation; however, their lateral veins are not numerous (only 6–9 pairs). Thus, the morphology of all three species is incompatible with diagnostic features of genus *Corylites*, and they should not be affiliated with this taxon.

Leaves of *C. amurensis* have been found together with bracteal involucre *Palaeocarpinus pacifica* sp. nov., and the absence of other leaves and generative organs of Betulaceae suggests that both kinds of remains represent the same plant. In addition, deposits of the Malomikhailovka Formation yielded rather numerous remains of coccuslike morphostructures 1–2 cm in diameter (Plate II, no. 11) most likely corresponding to *Ginkgo* seeds.

Distribution: Malomikhailovka Formation of the Amur River lower courses and Levaya Sobolevka Formation of the Sikhote Alin Ridge, Danian.

Material: more than 50 leaf impressions, e.g., specimens 143a, 258, 308b, 370, 385, 524, and others.

Palaeocarpinus pacifica Golovn. et Akhmetiev, sp. nov.
Plate II, no. 7; Fig. 3, no. 5

Holotype no. 3808/3c in collection of Geological Institute, Russian Academy of Sciences; Malomikhailovka Formation, Danian, Plate II, no. 7; Fig. 3, no. 5.

Description. The semiflosculous bracteal involucre is about 6 mm long and 4 mm wide, narrow at the base and rounded at the top, where it is decorated with several acute denses 1–1.5 mm long. In its basal part, there are several veins ramificating and anastomosing upward, where venation is of the parallel-dictyodromous type. Nutlets are missing.

Comparison and remarks. Infructescences of *Palaeocarpinus* were first described from the upper Paleocene deposits of southern England (Crane, 1981). They consist of small nutlets set in an envelope, which is composed of two deeply divided bracts equal in size. Similar fruits were later found in several localities of North America, France, and China, and at present, there are known to be several species of *Palaeocarpinus.* Bracts of the type species *P. laciniata* Crane from the Reading Beds are up to 8–15 mm long and divided into five lobes having large denses of irregular shape. Bracts of *P. joffrensis* Sun et Stockey from the Paleocene deposits of Canada are 3–6 mm long and have two or three lobes with denses varying in shape and dimensions (Sun and Stockey, 1992). As compared to both species, the bract from the Malomikhailovka Formation is not as lacinate and intermediate in size (larger than that of *P. joffrensis* and is lesser than in the case of *P. laciniata*). It is not divided into lobes, just having denses up to 1.5 mm long, and thus better resembling the undivided bracts of *P. joffrensis*, although denses of the latter are longer. Bracts of sole Asiatic species *P. orientalis* so far known only from northeastern China (Manchester and Guo, 1996) are very similar in morphology to those of *P. joffrensis.* Bracts of *Palaeocarpinus* affinity from the Paleocene Menat Beds of France have been described long ago under names *Corylus lamottei* Saporta (Saporta and Marion, 1885) and *"Atriplex" borealis* (Heer) Laurent (Laurent, 1912). They are up to 9 mm long, almost orbicular, and have margins divided into numerous acute denses. Thus, all *Palaeocarpinus* forms are highly variable in size, morphology, and marginal structure of their bracts. Since *P. pacifica* from our collection is represented by sole specimen, the diversity degree of its bracts, as well as size and morphology of its nutlets are so far unknown.

It is remarkable that leaves of Betulaceae, which have been found together with different *Palaeocarpinus* species, are also variable in morphology. For instance, *P. laciniata* coexists with narrow elliptical leaves of *Craspedodromophyllum acutum* Crane, which have the acute base and apex, and displays 12 to 16 pairs of secondary veins. Fruits of *P. joffrensis,*

Plate II. Danian plants from the Malomikhailovka locality (the Amur River lower courses). (1) *Amurocyparis sokoloviae,* shoot, specimen no. 77, ×1.5. (2) *Larix puerensis,* shortened shoots, specimen no. 382b, ×4. (3) *Larix olgensis,* specimen from herbarium of the Botanical Institute, Russian Academy of Sciences, ×1. (4) *Cryptomerites tutshkovii,* holotype, specimen no. 519, ×3. (5) *Corylites amurensis,* leaf margin, specimen no. 360, ×2. (6) *Larix puerensis,* leaf, paratype, specimen no. 3b, ×1. (7) *Palaeocarpinus pacifica,* bract, holotype, specimen no. 3c, ×7. (8–10) *Corylites amurensis,* specimens nos. 524 (8); 294 (9); and 385 (10), holotype. (11) *Ginkgo* sp. ?, seed, specimen no. 605, ×2.5.

P. orientalis, and *"Atriplex" borealis* occur together with predominantly oval leaves, which are cordate at the base and acute at the top. These leaves, which have 6–9 pairs of secondary veins, are most similar to leaves of *Corylus insignis* Heer from Canada, *C. macquarrii* (Forbes) Heer from France, and *C. bellatula* Guo et Li from China. In the Malomikhailovka Formation, *C. pacifica* is associated with elliptical to egg-shaped leaves of *Corylites amurensis* showing 14–19 pairs of secondary veins. Fruits of various *Palaeocarpinus* species and coexisting leaves of Betulaceae display a combination of features most often characteristic of representatives of Coryleae tribe (Sun and Stockey, 1992), but, in one way or another, all of its recent genera differ from this taxon. Taking into account the wide geographic distribution (the moderately warm zone of the Boreal belt) and diversity of vegetative organs of *Palaeocarpinus* forms, we believe that this genus was important for the Paleocene evolution of early Betulaceae.

Distribution: Malomikhailovka Formation, Amur River lower courses, Danian.

Material: specimen 3c.

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New Data on Eocene and Oligocene Phytostratigraphy of Severnye Ergeni (The South of the Russian Platform)

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Abstract—Assemblages of spores, pollen, and organic-walled phytoplankton were studied in the Middle Eocene–Oligocene sequence of Severnye Ergeni. The dinocyst zonal assemblages were described, and anomalies in their composition related to the temperature regime and hydrology of the sea basin were outlined. Anoxic environments in the Bartonian and Chattian time were recognized. The considerable role of the *Adnatosphaeridium–Glaphyrocysta–Cyclonephelium* group indirectly implies a warmer state and freshening of the sea basin of the Solenaya Balka time interval; a warmer climate is also suggested by presence of various thermophilic taxa in pollen assemblages and by the greater amount of pollen of Fagaceae and subtropical plants.

Key words: *Dinocysts, prasinophytes, acritarchs, spores, pollen, Eocene, Oligocene, horizon, assemblage, paleoclimate.*

INTRODUCTION

The village of Ergeni is located in the interfluve area of the lower reaches of the Volga and Don rivers and is considered to be one of the classical regions of deposits of the second half of the Paleogene in the Russian platform. Though the number of exposures in river valleys is limited, their sections were studied in detail and intercorrelated by using data on a dense network of cartographic boreholes drilled for hydrotechnical purposes and also for prospecting the Paleogene beds. Particularly favorable for this study were the Maikop deposits accumulated in the outer shelf zone of an epicontinental sea; they are several hundreds of meters thick and differ from the inner beds of the Maikop basin in having an abundance of different assemblages of organic remains. It was this region where G.I. Garetskii in 1953 identified the Solenaya Balka Horizon, one of the most easily identifiable and persistent reference levels in the Maikop series. Precisely this horizon yielded the typical brackish-water mollusk *Ergenica cimlanica* (Popov) and the endemic ostracod assemblage, which allowed us to reliably correlate the level of the Solenaya Balka deposits in Hungary and Romania in the west to the Aral area in the east. Until now, however, an adequate biostratigraphic study of Paleogene sections in the region was not conducted, and attention was focused solely on benthic foraminifers recovered from drilling cores. Though these organisms provided a highly detailed subdivision of the section, their composition often largely depends on the facies environment.

The objective of this study was to investigate the section of Borehole 45A, penetrated through the 300-m-thick Paleogene deposits. The borehole was drilled in Severnye Ergeni, 12 km northeast of the Malye Derbety settlement (Fig. 1). The obtained data was supplemented by results of a study of Maikop deposits recovered from the borehole drilled near Tsatsa Lake, north of Borehole 45A. My task was to study spores, pollen, and organic-walled phytoplankton (the dinoflagellate cysts, green algae, and acritarchs) and to define the age of separate intervals of the section on the basis of these

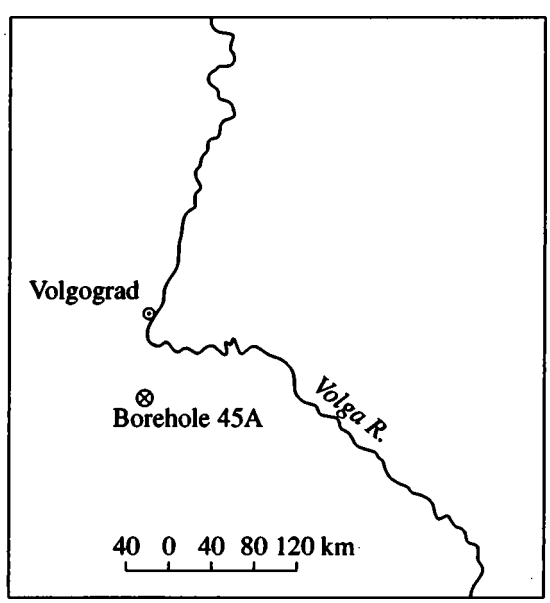


Fig. 1. Location of Borehole 45A in Severnye Ergeni (lower reaches of the Volga River).

organic remains and their correlation with data on malacofauna, foraminifers, and nannoplankton that were obtained by S.V. Popov, V.N. Ben'yamovskii and N.G. Muzylev (personal communications) from the same section.

A GENERAL DESCRIPTION OF THE SECTION

In Borehole 45A, the Paleogene deposits were penetrated in the interval of 362–48 m. Above, the Oligocene beds are covered by the Pliocene Ergeni sands.

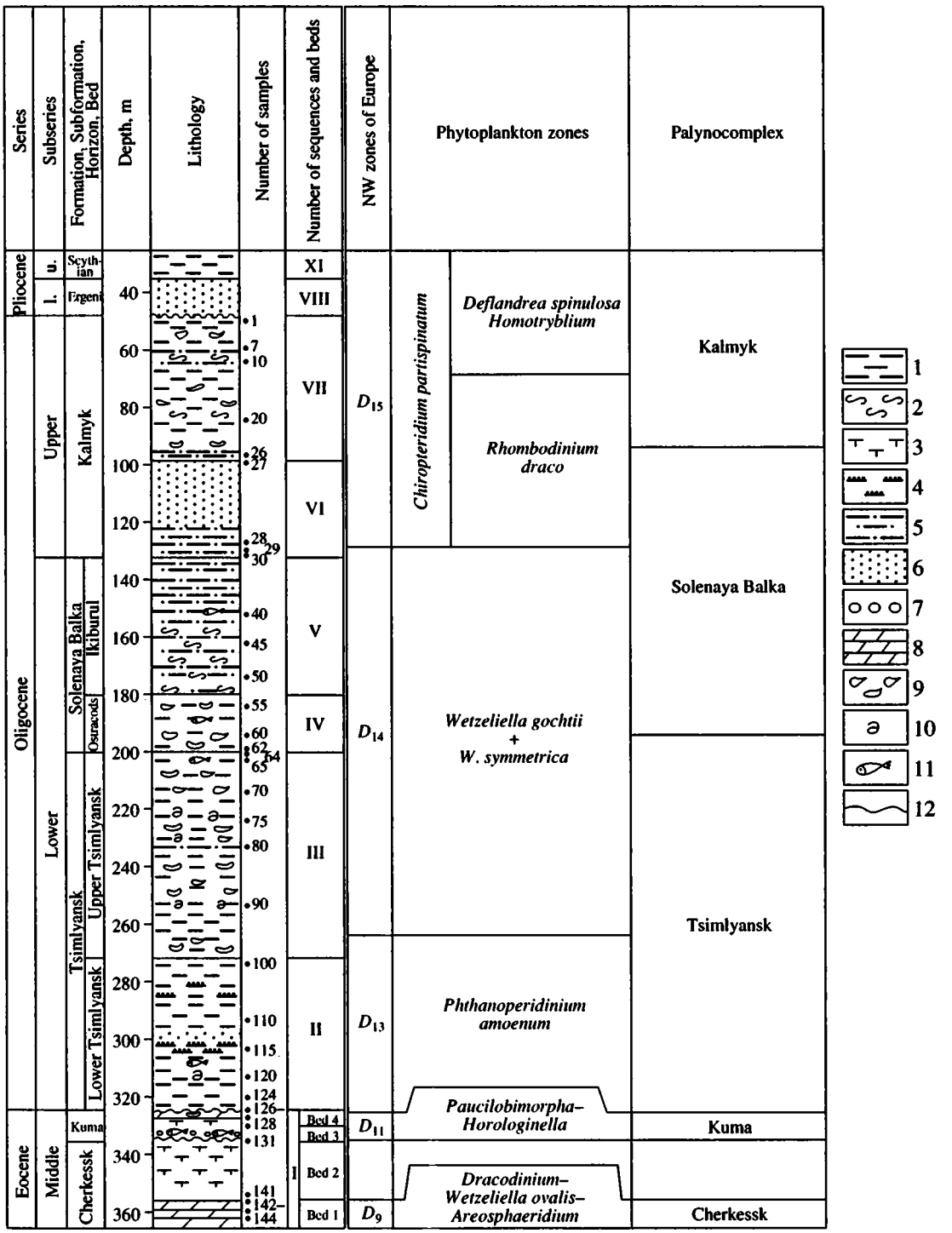


Fig. 2. Phytoplankton zones and pollen assemblages of the Eocene and Oligocene in the section of Borehole 45A: (1) clays; (2) noncarbonate clay; (3) carbonate clay; (4) clays with silt fraction on bedding planes; (5) silt; (6) sand; (7) shingle with phosphorite nodules; (8) marl; (9) nodules of ferruginous carbonate, marcasite, and pyrite; (10) rounded clayey pellets; (11) fish remains; (12) erosion boundary.

The entire Paleogene section is characterized by 32 studied samples, 24 of which belong to its Oligocene part and 8 to the Eocene beds. Unfortunately, two important intervals of the Eocene and Oligocene were not sampled (357–337 and 125–100 m). Since nearly all samples contain abundant palynomorphs, they are suitable for statistical processing, and the number of calculated grains in a sample varied from 350–400 to more than 1500. In accordance with the initial documentation, the whole Paleogene section was divided into a series of beds representing individual units of local stratigraphic subdivisions (Fig. 2).

The Middle Eocene[1]

Cherkessk Formation. The upper part of the formation is referred to the Keresta Horizon. Member I (362–334 m).

Bed 1. White marls.

Bed 2. Dark gray carbonate clays.

Hiatus.

Kuma Horizon. Member I (334–325 m).

Bed 3. Carbonate marls and clays with fish remains. At the base, there are phosphorite concretions and gravel up to 1 cm in diameter.

Bed 4. Greenish gray marls.

Hiatus.

Lower Oligocene. Lower Maikop

Tsimlyansk Formation. Lower Tsimlyansk Subformation. Member II (325–272 m).

In the lower part of the member, there are fine-laminated dark gray clays with carbonate interbeds and rounded clayey pellets. In the middle part, the fine-laminated dark gray clay with micaceous silt interbeds (1–2 cm) contains abundant fish scales. In the upper part, the clay is almost black and includes aleurite powder on the upper planes of individual laminae.

Upper Tsimlyansk Subformation. Member III (272–203 m).

In its lower and middle parts, the member consists of dark gray clay with aleuritic interbeds displaying an obscure lamination. In the upper part, the clay contains greenish gray pyrite concretions.

Solenaya Balka Horizon. Ostracod Beds. Member IV (203–180 m).

Aleuritic, massive, bluish-gray clay with interbeds of marls and limestones (?).

Ikiburul Beds. Member V (180–132 m).

Noncarbonate and low-carbonate clays, clayey silts, dark gray and laminated.

[1] Leonov (1957) considered this part of the section on the right-hand bank of the Volga to belong to the Aksai and Balyklei beds and to be the late Eocene in age.

Upper Oligocene. Middle Maikop

Kalmyk Formation. Member VI (132–98 m).

At the base, there are sandy silts, micaceous, gray. Upsection they grade into fine-grained sands ferruginate at the base.

Member VII (98–48 m).

Above the basal white aleurites with sand intercalations, a larger part of the member is composed of interbedded clays and silts with melnikovite nodules and mud-eater burrows. In the upper part of the member, the clayey bluish gray silts are overlain by a thin clay bed with marly bluish to greenish gray concretions.

Upsection, the Pliocene Ergeni sands (48–27 m) are overlain with a hiatus by the Scythian brick-red clays (27–0 m).

In this section, relations between the organic-walled phytoplankton, spores, and pollen are variable (Figs. 3, 4). In the Eocene deposits, phytoplankton clearly prevails; in the Oligocene rocks it usually represents from 5 to 25% of the total amount of palinomorphs. In the marine facies of the Tsimlyansk and Kalmyk formations, the organic-walled phytoplankton is more diverse and includes, in some samples, up to 50 taxons and more. The abundance of phytoplankton is reduced in brackish-water ostracod and Ikuburul beds. This interval is also characterized by a high concentration of amorphic organic matter in preparations. The quantitative relationships of the different organic-walled phytoplankton groups vary along the section. Representatives of the *Paucilobimorpha* genus of green algae visibly dominate in the Kuma Horizon, whereas acritarchs *Micrhystridium* dominate in the Tsimlyansk Formation, particularly near the boundary with the Solenaya Balka Horizon. The composition of acritarchs and green algae is also diverse in basal beds of the Kalmyk Formation, directly below the 20-m-thick sand member that is unsampled. In the other intervals of the section, pyrrhophyte algae dominate in the phytoplankton spectrum.

PALYNOLOGICAL CHARACTERISTICS

In the Eocene part of the section, two assemblages different in composition from the main palynomorphs are identified.

The marl layer at the base of the visible section of the Cherkessk Formation (362–356 m) yielded the phytoplankton assemblage of *Dracodinium–Wetzeliella ovalis–Areosphaeridium* type, where the species of the first two genera are dominant (*Dracodinium simile, D. pachydermum, D. varielongitudum*, and also *Microdinium reticulatum, M. ornatum*). They are accompanied by species (sometimes abundant) of genera *Spiniferites* (*S. membranophoridium, S. cornutus, S. pseudofurcatum*), *Cordosphaeridium* (*C. gracile, C. inoides, C. biarmatum*), and *Areosphaeridium* (*A. arcuatum, A. diktyoplokus*). In this interval, up to 50 taxons were found as single specimens: *Hystrichostrogylon membranophoridium,*

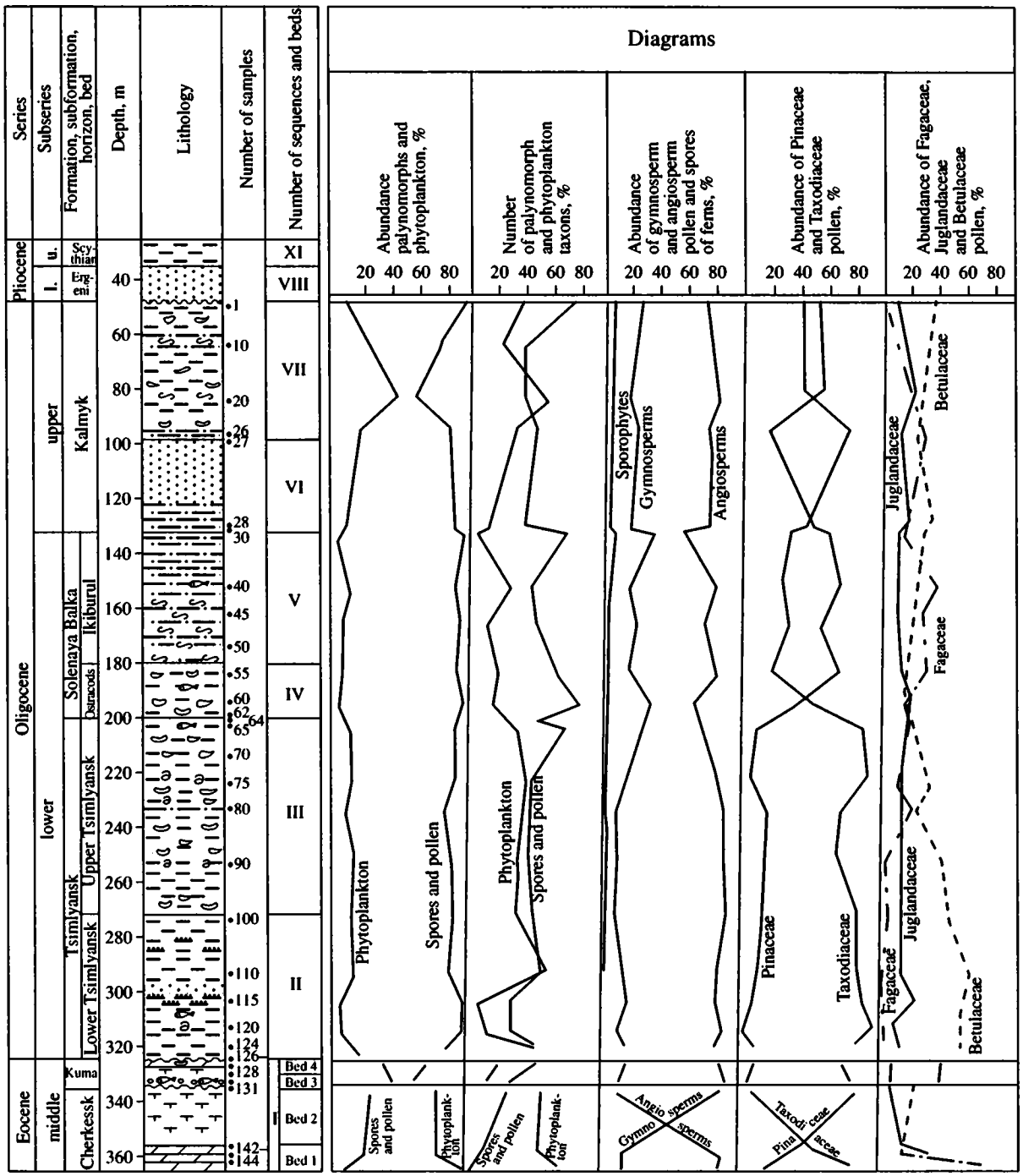


Fig. 3. Palynodiagrams for Eocene and Oligocene deposits penetrated by Borehole 45A (see explanations nos. 1–19 to Fig. 2).

H. membranophoridium subsp. *granulatum, Diphyes ficusoides, D. colligerum, Corrudinium, Tectatodinium, Thalassiphora elongata, T. pelagica, T. delicata, Wilsonidium echinosuturatum, Deflandrea dissoluta, D. oebisfeldensis, D. leptalea, Systematophora placacantha, Areoligera, Dapsilidinium, Melitasphaeridium asterum, Operculodinium* (including *O. nanaconulum*), *Achilleodinium biformoides, Alterbidinium, Hystrichokolpoma cinctum, H. salacium, Heteraulacacysta leptalea, H. porosa, H. campanula, Rottnestia borussica, Lingulodinium, Eatonicysta ursulae* subsp. *furensis, Laternosphaeridium* (cf. *L. lanosum*), *Dinopterygium cladoides* sensu Morgenroth, *Apectodinium* cf. *Wetzeliella echinulata, Oligosphaeridium, Cleistosphaeridium, Cerodinium, Alisocysta, Melliodinium, Impagidinium* (including *I. brevisulcatum*), *Fromea, Achomo-*

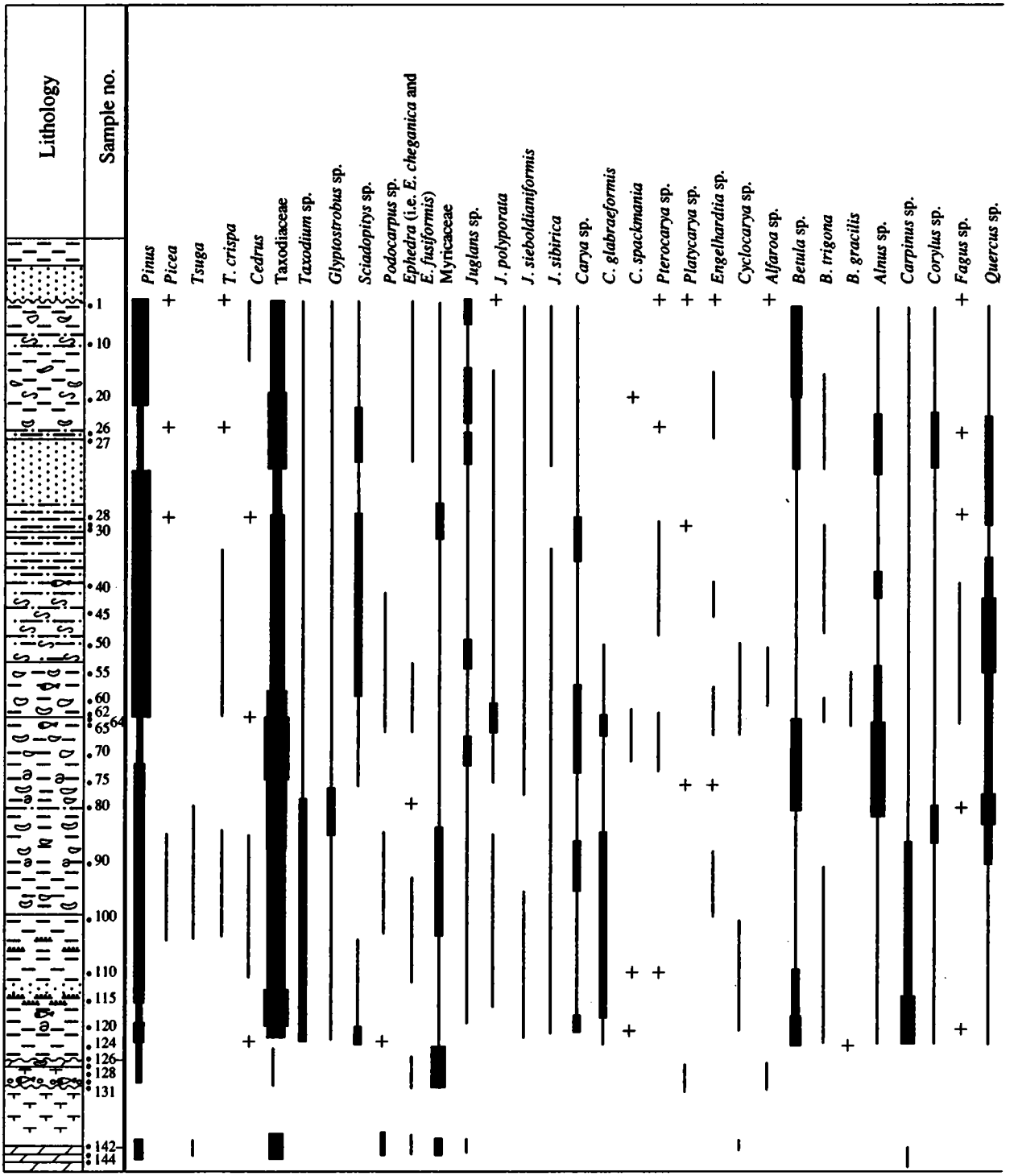
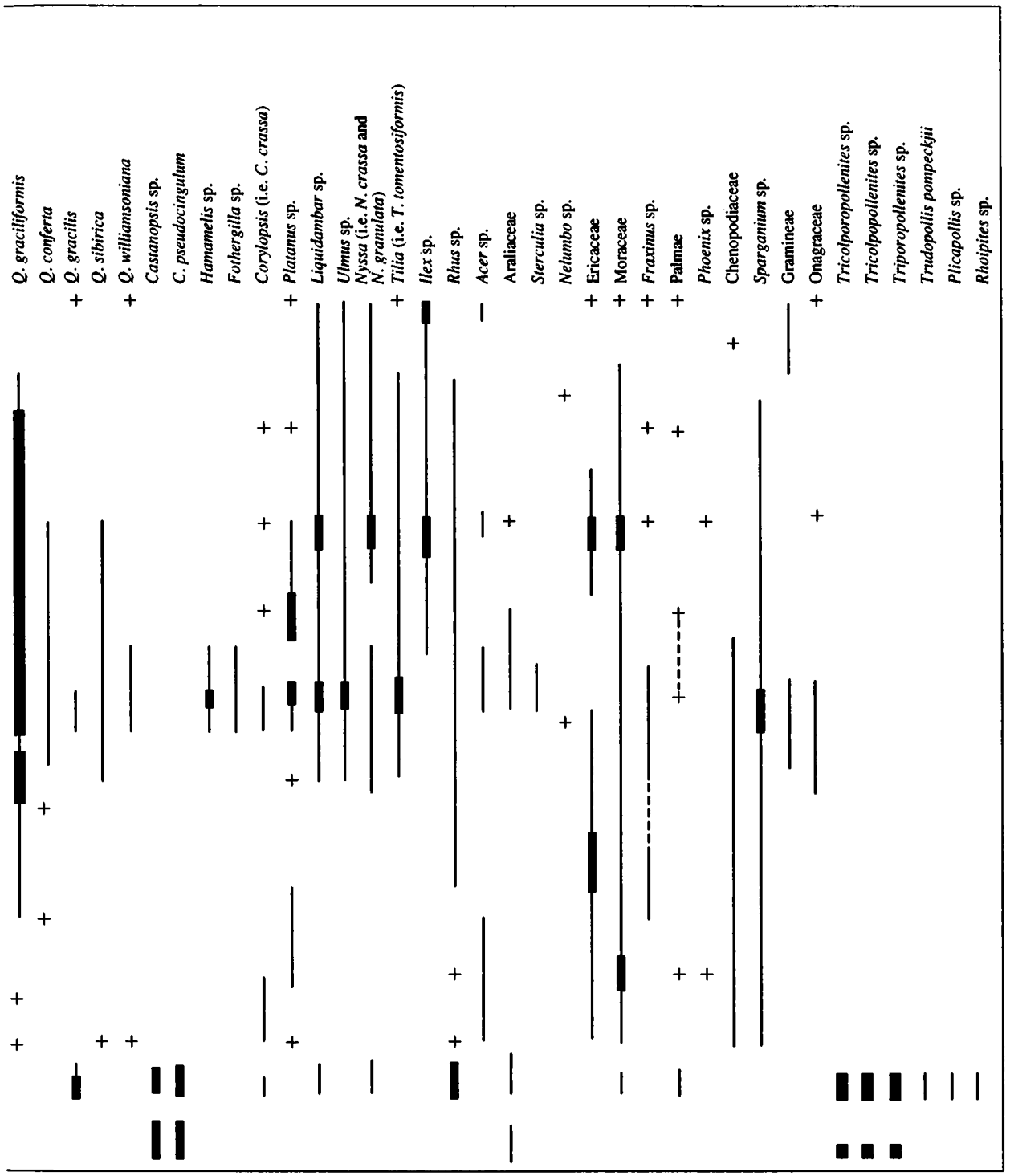


Fig. 3. (Contd.)

sphaera alcicornu, Cerebrocysta, Phthanoperidinium, Samlandia chlamidophora, and *Pentadinium taeniagerum.* The prasinophytes and acritarchs group includes single specimens of *Micrhystridium.*

The joint occurrence of the *Dracodinium* species, typical of the lower Eocene and Lutetian, with *Areosphaeridium* and *Wetzeliella ovalis* species, which appear in the middle Eocene, indicates the Lutetian rather than the Bartonian age for this part of the section. Muzylev refers it, on the basis of nannoplankton, to the upper Lutetian and Ben'yamovskii and on the basis of foraminifers, to the lower Bartonian (personal communication).



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The pollen assemblage in marls is poor and includes pollen of gymnosperms, mainly Taxodiaceae, as the dominant component associated with accessory pollen of *Podocarpus* and *Ephedra*. The angiosperms are represented by rare pollen grains of Amentoflorae *Myrica*, *Cyclocarya*, and *Castanopsis pseudocingulum*. There are also the tricoplate and tricoplate–cribellate pollens of the formal taxons.

The phytoplankton from clayey–marl beds of the Kuma Horizon (336–325 m) belongs to *Paucilobimorpha–Horologinella* assemblage, in which these index taxons constitute 50% of palynomorphs. The main compo-

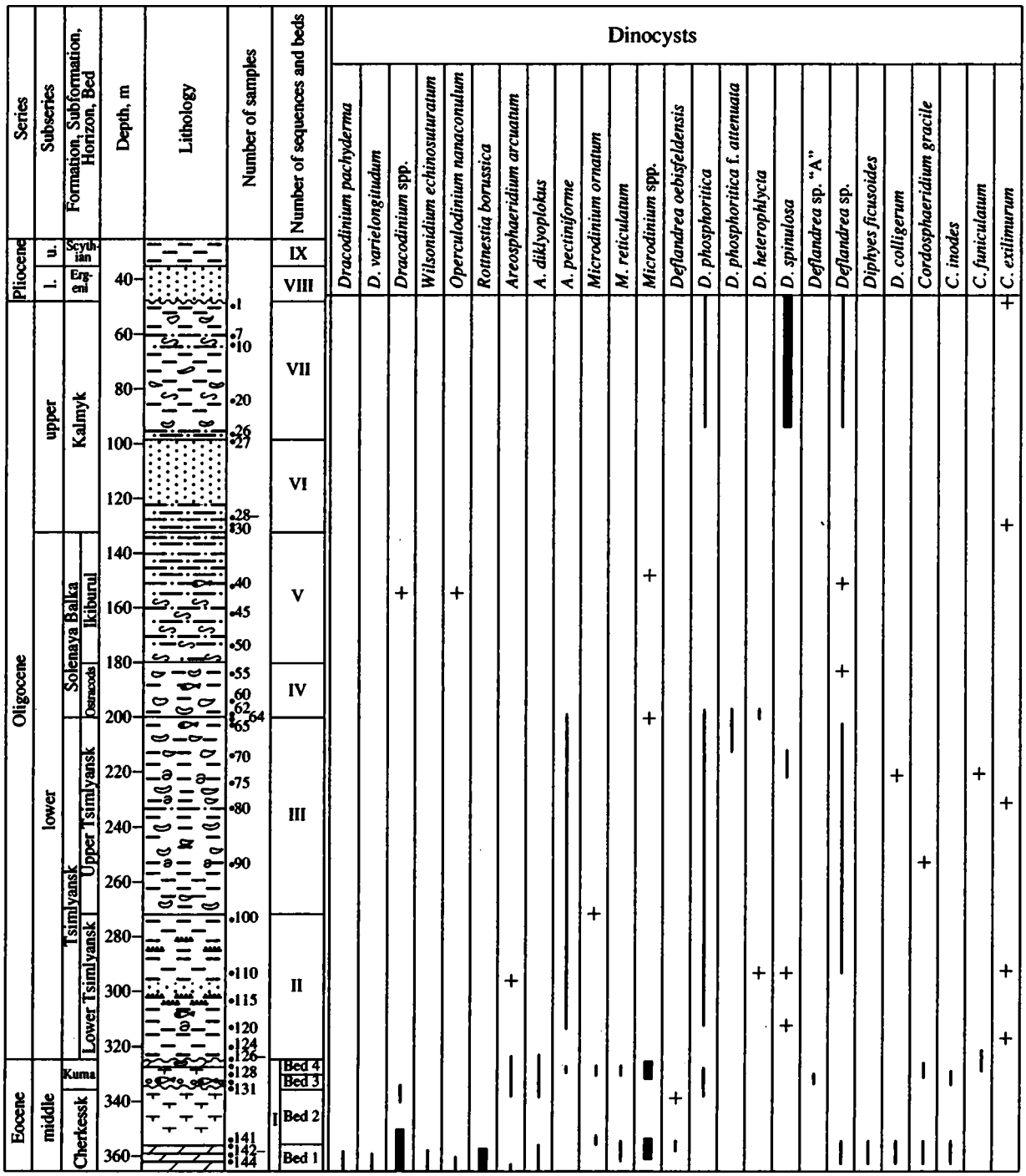
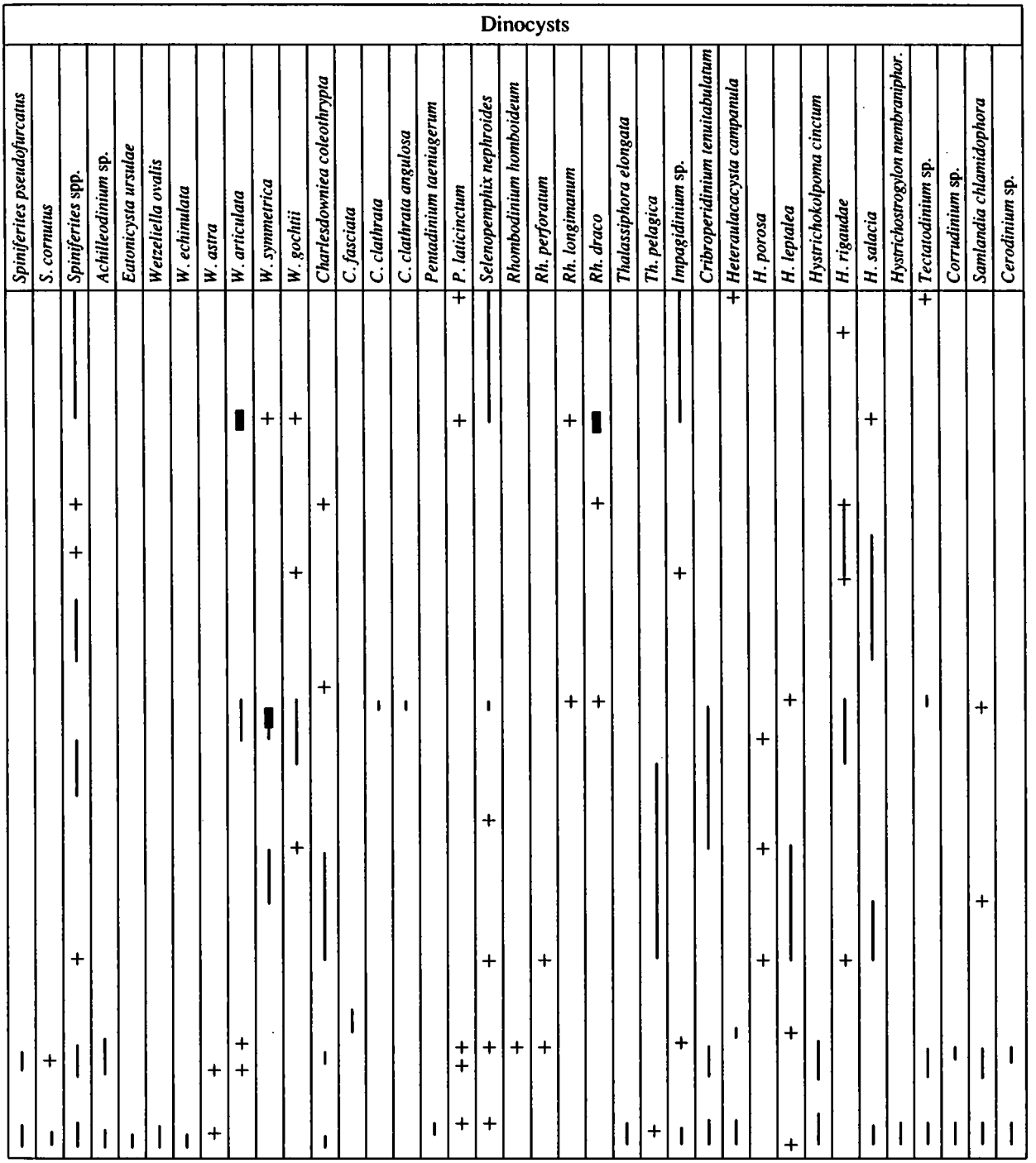


Fig. 4. Distribution of dinocysts, green algae, and acritarchs in Borehole 45A (see explanations nos. 1–19 to Fig. 2).

nent is *P. triradiata* (more than 120 specimens per slide), and several *P. granulifera* species of the same genus. Among dinocysts, the most important are species of genera *Areosphaeridium* (*A. arcuatum, A. diktyoplokus, A. pectiniforme*), *Deflandrea* (*D. arcuata, D. phosphoritica, D. oebisfeldensis, Deflandrea* sp. A) and *Microdinium* (*M. ornatum, M. reticulatum*). Representatives of no fewer than 50 dinocyst taxons were recovered as rare specimens, including *Apteodinium* cf. *A. maculatum, Spiniferites* (*S. ramosus, S. cornutus, S. pseudofurcatus*), *Charlesdowniea coleothrypta, Wetzeliella articulata, W. astra, Dracodinium* cf. *D. condylos, Apectodinium longispinosum, A.* cf. *homomorphum, Heteraulacacysta pustulata, Lingulodinium,*



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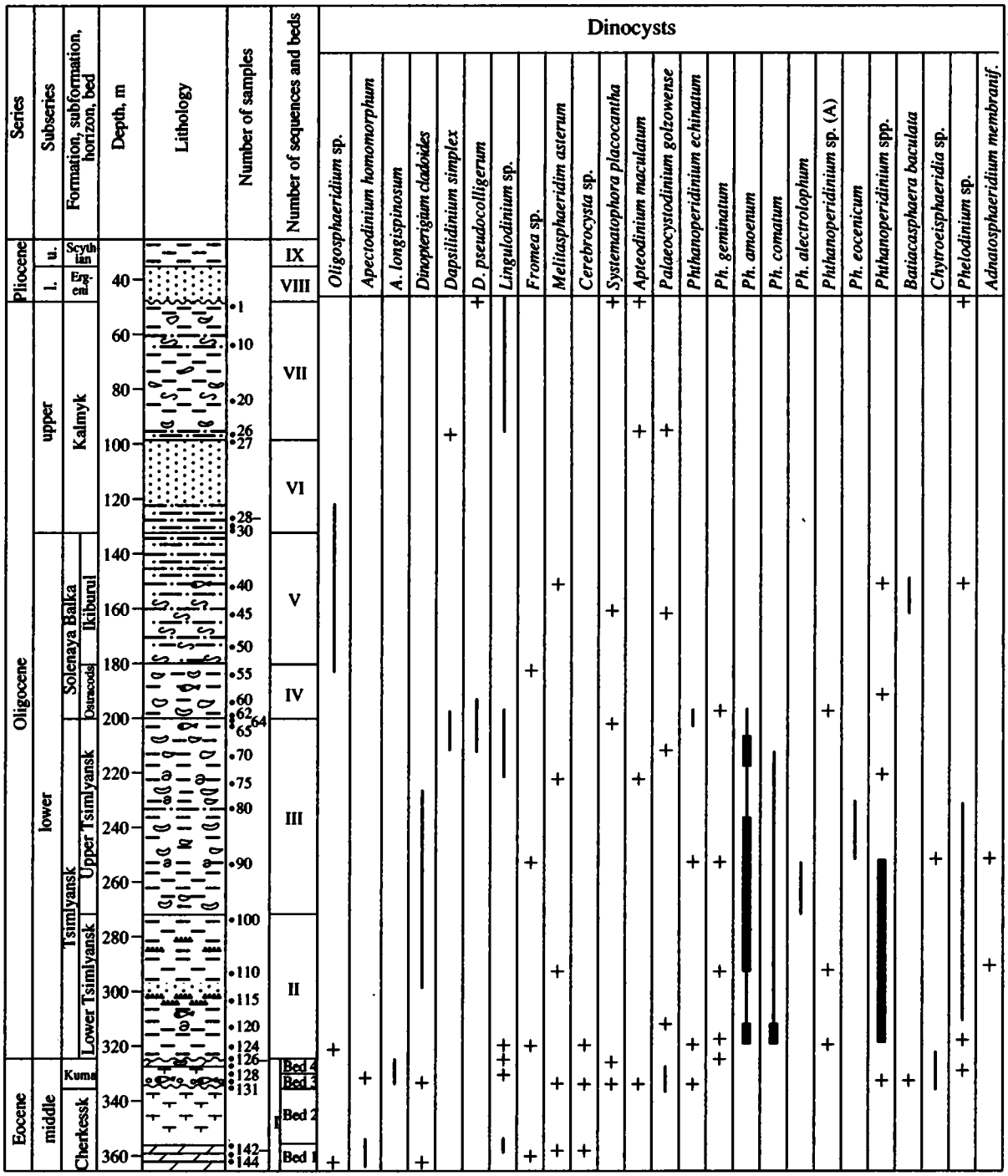
Samlandia chlamidophora, Cordosphaeridium gracile, C. inodes, Melitasphaeridium, Spinidinium, Heteraulacacysta, Impagidinium, Rotnestia borussica, Hystrichosphaeropsis, Hystrichostrogylon, and *Fibrocysta.* Acritarchs and prasinophytes are represented by *Ptanospermella* and *Micrhystridium.*

The pollen composition in the upper horizons of the Cherkessk Formation differs from that in the Kuma Horizon. The pollen of gymnosperms, mainly of *Pinus,* and rare grains of Taxodiaceae and *Ephedra* constitute no more than 10% of the assemblage, the bulk of which consists of Fagaceae pollen represented by fine pollen of *Quercus gracilis, Q. graciliformis* and *Q. conferta* usual for the Eocene. Samples 131 and 128 yielded pollen of *Castanopsis* in detectible quantities. The amount of pollen of formal taxons is high (up to 30%); they



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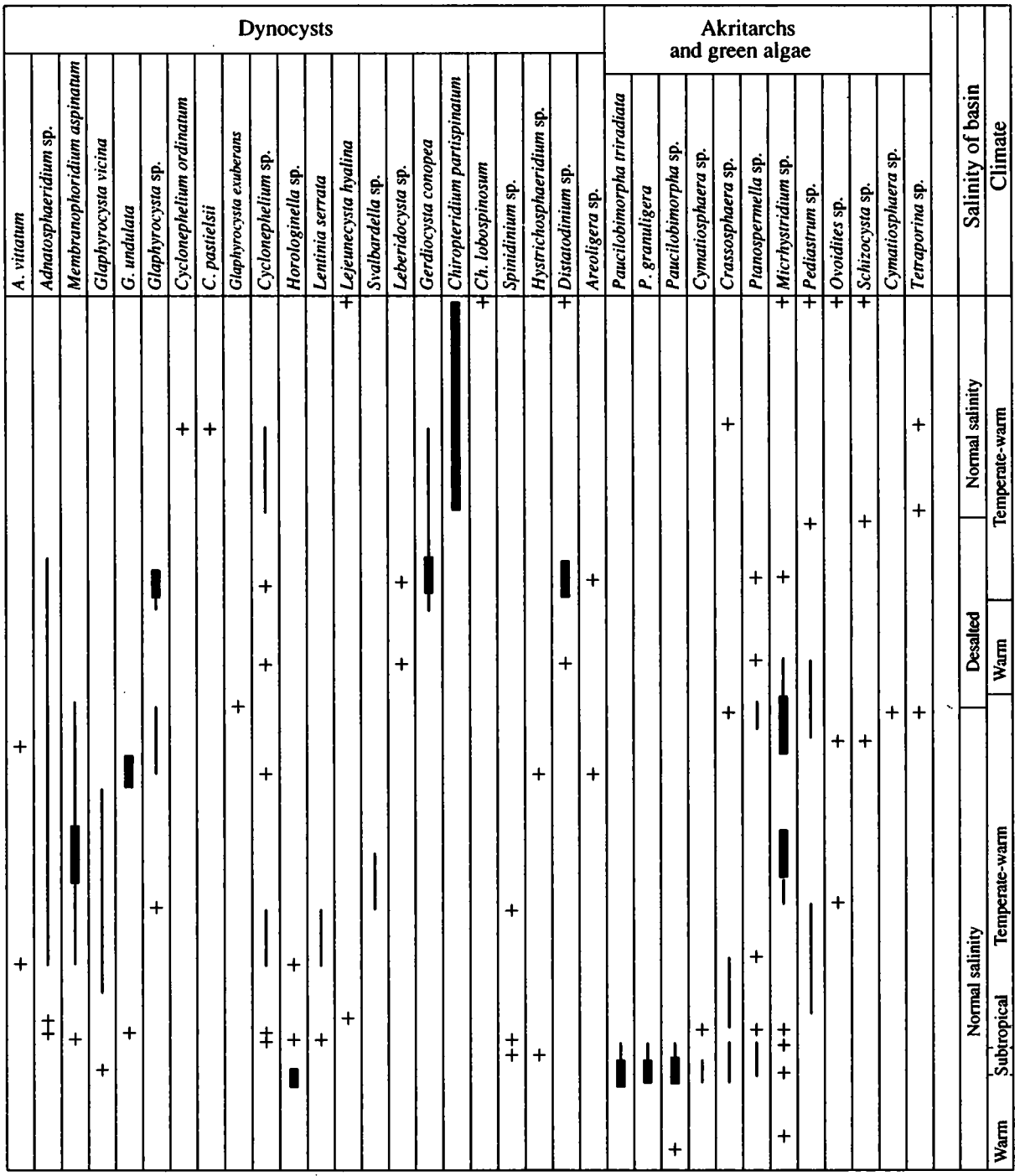
include *Plicapollis, Triporopollenites* cf. *T. megagranifera, Tricolporopollenites, Triporopollenites, Trudopollis pompeckjii, Rhoipites,* and *Triotriapollenites.* The pollen of Juglandaceae is usually represented by thermophilic *Alfaroa* and *Engelhardtia* forms, in association with the pollen of Palmae, Araliaceae, *Comptonia minima, Rhus, Corylopsis,* and *Liquidambar.*

The age of the Kuma Horizon was determined with due consideration of the fact that, in the Paleogene sections of western Europe (the southern Netherlands, Great Britain), species *Paucilobimorpha* and *Horologinella* are quite common in the upper Lutetian, particularly in the Bartonian deposits (De Coninck, 1986). Many phytoplankton species characteristic of the dinocyst zone *Rhombodinium perforatum–R. porosum*



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also suggest that this horizon is attributable of the second half of the Bartonian. This conclusion is confirmed by data on foraminifers and nannoplankton from the same beds.

The Eocene interval of the section shows a wide spectrum of dinoflagellate cysts indicating, an open warm sea basin of normal salinity that stretched from the west and south into the area of the southeastern margin of the Russian platform. The Bartonian seasonal climate was of the subtropical type, though less humid and with warmer summer periods than those of the Lutetian time.

The upper part of the studied Eocene section (sample 126) with numerous redeposited forms of the Paleo-

gene and Cretaceous represents a horizon of reworked deposits. According to evidence detected in the section of the Kuma Horizon, the respective hiatus in sedimentation was apparently brief.

In the Oligocene part of the section, data on phytoplankton and palynomorphs of higher plants revealed a regular succession of assemblages. Judging by zonal phytoplankton species, this part includes all zones of the dinocyst scale adopted for the Crimea–Caucasus region and the Russian platform.

Three zonal assemblages of the lower Oligocene (*Phthanoperidinium amoenum, Wetzeliella symmetrica,* and *W. gochtii*) were identified in the Tsimlyansk Formation. The microflora of the Solenaya Balka Horizon is referred to the third assemblage. If the lower–upper Oligocene boundary separates *Wetzeliella gochtii* and *Chiropteridium partispinatum–Rhombodinium draco* zones, then it lies within the boundary interval between members V and VI, which separates the Ikiburul Beds of the Solenaya Balka Horizon and the Kalmyk Formation. The following consideration, however, should be taken into account: phytoplankton assemblages are insufficiently diverse in all samples below the base of the sandy member in the lower part of the Kalmyk Formation. They mainly consist of green algae and acritarchs. Their host rocks are rich in amorphic organic matter, which implies an anoxic environment of the basin retained from the Solenaya Balka time. Enviroments with normal water salinity existed after, or during, the period of sand accumulation in the course of maximal regression and restricted contact with the open ocean.

Before the further description of phytoplankton assemblages, it is necessary to allocate ranges of their zones in the Paleogene scale.

Phthanoperidinium amoenum Zone (325–275 m) is the lower unit of the Oligocene corresponding to the lower part of the P18 (*Globigerina tapuriensis*) Zone of the planktonic foraminifer zonation and to the NP21 and NP22 zones (the lower part) of the nannoplankton zonation. This interval of both scales corresponds to the lower part of the Rupelian Stage. In the Landzhar section (Armenia), the Eocene–Oligocene boundary stratotype for the territory of the former USSR, the zonal species *Ph. amoenum* first appears at the base of the P18 Zone. In the section of Borehole 45A, it is found in abundance in samples 110, 100, and 90 from the depth interval of 294–254 m, i.e., from the very base of the Maikop Group.

The undivided *Wetzeliella symmetrica–W. gochtii* Zone was identified in the depth interval of 275–132 m. Its upper boundary is drawn at the first occurrence of *Chiropteridium partispinatum* and is adopted as the boundary between the upper and lower Oligocene, though many researchers, e.g., Andreeva-Grigorovich (1991) and others, consider the *Ch. partispinatum* zone as a part of the lower Oligocene. The *Wetzeliella symmetrica–W. gochtii* Zone is undivided because of difficulties in classifying wetzeliellas at the species rank, owing to their poor preservation. It can be tentatively suggested that the *W. symmetrica* Zone covers the interval of 275–225 m, whereas the upper horizons of the upper Tsymlyansk Subformation are reliably referred to the *W. gochtii* Zone. In the dinocyst zonation of Western Europe, the *W. symmetrica* Zone is correlated with the NP22 Zone, while *W. gochtii* is considered as compatible with NP23 Zone of the nannoplankton scale. According to planktonic foraminifer zonation, the lower unit corresponds to the *Globigerina tapuriensis* Zone, and the *W. gochtii* Zone, though only partly, is placed at the level of the *Globigerina sellii* Zone. In Borehole 45A, *W. symmetrica* appears as rare specimens in samples 100, 90, 75, and 65. Both zonal species are present in sample 70, and *W. gotchii* are detected in samples 75 and 64. Both species also appear episodically in the Solenaya Balka Horizon and even in lower beds of the Kalmyk Formation.

The *Ch. partispinatum* Zone covers the rest of the drilled Oligocene section above the Solenaya Balka Horizon. In the dinocyst zonation, its range is treated differently. J. Chateauneuf, as mentioned in Cavelier and Pomerol (1983), believes that this zone characterizes the lower Oligocene below its upper part, and he places it above two other zones marked by *Lingulodinium* (*Cordosphaeridium*) *xanthium* and *Homotryblium floripes.* Andreeva-Grigorovich (1991), in her analysis of dinocyst zonation accepted for the southern regions of the former USSR, considerably extends the range of this zone, including in it the entire upper and a part of the lower Oligocene. She subdivided the zone in the North Caucasus sections into beds with *Chiropteridium partispinatum* s.s., *L. xanthium* and *H. floripes.* In the section of Borehole 45A, the *Ch. partispinatum* Zone s.l. can be also divided into two subzones or beds: one with *Ch. partispinatum–Rhombodinium draco* assemblage (samples 28, 27, 26) and another with *Ch. partispinatum–Deflandrea spinulosa* and *Homotryblium* ex gr. *floripes–H. tenuispinosum* assenblages. As regards the *Lingulodinium* genus, it is found in all the samples collected from the Kalmyk Formation.

Below, the composition of the zonal assemblages is given in greater detail (Fig. 2).

Assemblage with Phthanoperidinium amoenum. In addition to zonal species, the assemblage includes other forms of this genus, i.e., *Ph. comatum, Ph. geminatum, Phthanoperidinium* sp. A (with echinulate surface), *Ph. echinatum, Ph. alectrolophum,* ? *Ph. eocenicum,* and rare specimens of *Areosphaeridium pectiniforme, Homotryblium abbreviatum,* and the *Membranophoridium–Glaphyrocysta–Cyclonephelium* group, including *M. aspinatum, G. semitecta,* the *G. vicina* typical of the first half of the early Oligocene.

In the upper part of the Tsimlyansk Subformation, the phytoplankton is more diverse (>50 taxons) and characterizes marine facies of the lower part of the Oligocene section. Phytoplankton makes up 20–25% of

the total amount of palinomorphs and contains many taxons, such as *Deflandrea, Charlesdowniea, Heteraulacacysta* (including *H. porosa* and *H. leptalea*), *Rhombodinium, Adnatosphaeridium* (*A. vittatum, A. multispinosum*), *Hystrichokolpoma* (including *H. rigaudae* and *H. salacium*), Thalassiphora (*Th. pelagica* and *Th. reticulata*), *Cordosphaeridium exilimurum, Lentinia, Samlandia chlamidophora, Operculodinium, Phelodinium, Spiniferites, Pentadinium, Dinopterygium cladoides* sensu Morgenroth, *Microdinium ornatum, Spinidium, Svalbardella, Cribroperidinium tenuitabulatum, Fromea,* and others.

The assemblage of undivided *Wetzeliella symmetrica–Wetzeliella gochtii* Zone is typical of the upper Tsimlyansk Subformation and Solenaya Balka Horizon. The diversity of phytoplankton species suddenly changes starting from the base of the ostracod beds, and mollusks *Ergenica cimlanica* begin to appear from that level. From depths of 254 m (sample 90) and higher, including the base of the ostracod beds, there are many *Micrhystridium* forms, though not in all samples, which is an indication of the coastal facies deposited under notable influence of fresh water. In the upper part of Tsimlyansk Formation, which was formed under normal water salinity, the phytoplnkton composition in the depth interval of 275–203-m, is as follows.

(1) *Deflandrea* species: *Deflandrea phosphoritica* (with subspecies *D. ph. phosphoritica* and *D. ph. attenuata*), *D. heterophlycta,* (?) *D. spinulosa,* and *D. andromensis.*

(2) Species of the generic group *Wetzeliella–Charlesdowniea: Wetzeliella gochtii, W. symmetrica, Wetzeliella* f. A, *W. articulata, Charlesdowniea clathrata,* and *Ch. clathrata angulosa.*

(3) *Phthanoperidinium* species: *Phthanoperidinium amoenum, Ph. echinatum,* and *Ph. comatum.*

In this assemblage, there are also *Homotryblium floripes, H. plectilum, H. abbreviatum, Dapsilidinium pseudocolligerum, D. simplex, Glaphyrocysta exuberans, Membranophoridium aspinatum, Cyclonephelium, Adnatosphaeridium vittatum, Cribroperidinium tenuitabulatum, Systematophora placantha, Dapsilidinium simplex, D. pseudocolligerum, Heteraulacacysta porosa, Hystrichokolpoma rigaudiae, Areosphaeridium pectiniforme, Lingulodinium machoerophorum, Spiniferites ramosus, Cordosphaeridium* sp., *Apteodinium* sp., *Rhombodinium* sp., *Adnatosphaeridium* sp., *Tectatodinium* sp., *Microdinium* sp., and *Cordosphaeridium fibrospinosum.*

Green algae and the acritarch group are represented by *Micrhystridium* and by rare specimens of *Crassosphaera, Tetraporina, Cymatiosphaera, Ptanospermella, Pediastrum, Schizocysta,* and *Ovoidites.*

In sample 64 from the basal part of the sequence with *Ergenica cimlanica* (202-m depth), the phytoplankton composition is the same.

The phytoplankton assemblage from the Solenaya Balka Horizon is very poor, which is typical for brackish-water deposits (abundance less than 5% of the total number of palinomorphs).

Sample 60, recovered from the depth of 194 m, contains *Adnatosphaeridium* sp., *Hystrichokolpoma salacia, Gerdiocysta, Batiacasphaera, Homotryblium floripes, Dapsilidinium, Phthanoperidinium, Charlesdowniea clathrata,* and *Apteodinium emslandense.*

Sample 55 from the depth of 184 m contains species ex. gr. *Spiniferites, Adnatosphaeridium* sp., ex gr. *Cyclonephelium,* ex gr. *Dystatodinium, Oligosphaeridium* sp., *Leberidocysta,* cf. *Hapsocysta, Fromea, Hystrichokolpoma salacia, Daflandrea, Homotryblium plectilum, Dapsilidinium, Phthanoperidinium.*

Sample 50, recovered from the depth of 174 m, contains *Hystrichokolpoma salacia* (>10%), *Hystrichokolpoma* sp., *Distatodinium, Oligosphaeridium, Adnatosphaeridium, Gerdiocysta, Cyclonephelium reductum,* (all up to 5%), and rare specimens of *Oligosphaeridium complex, Melitasphaeridium, Impagidinium, Kallosphaeridium, Batiacasphaera, Apteodinium, Glaphyrocysta, Chytroeisphaeridia,* cf. *Gonyaulacysta diutina, Heteraulacacysta, Cleistosphaeridium, Rhombodinium,* and *Spiniferites. Pediastrum* forms dominate among prasinophytes and acritarchs, which are represented by single specimens of *Micrhystridium* and *Ovoidites.*

Sample 45, recovered from a depth of 164 m, contains *Palaeocystodinium golzovense, Hystrichokolpoma salacia, Phthanoperidinium* sp., *Homotryblium,* cf. *Oligosphaeridium, Melitasphaeridium,* ex gr. *Adnatosphaeridium,* ex gr. *Spiniferites, Wetzeliella, Charlesdowniea, Batiacasphaera baculata, Glaphyrocysta,* and *Gerdiocysta conopea.*

Sample 40, from a depth of 154 m, yielded *Glaphyrocysta* spp. (numerous specimens), *Adnatosphaeridium, Gerdiocysta, Phelodinium, Phthanoperidinium, Microdinium, Wetzeliella gochtii, Wetzeliella* f. A, *Distatodinium, Impagidinium, Homotryblium, Batiacasphaera, Deflandrea, Oligosphaeridium, Dracodinium, Cordosphaeridium, Melitasphaeridium,* cf. *Leberidocysta,* and *Systematophora.* The group of prasinophytes and acritarchs consists of *Ptanospermella* forms.

Sample 30, from a depth of 133 m, contains *Gerdiocysta, Cordosphaeridium exilimurum,* and *Operculodinium.* The group of prasinophytes and acritarchs includes *Schizocysta, Micrhystridium,* and *Ptanospermella.*

Listed taxons show that the *Glaphyrocysta–Cyclonephelium–Adnatosphaeridium* group and rare specimens of *Gerdiocysta, Batiacasphaera,* and *Leberidocysta* absent in other intervals are characteristic of brackish-water deposits. The last taxon is rather typical of the Solenaya Balka Horizon in northern Caucasus (Zaporozhets, 1993).

Sample 29, from a depth of 130 m, showed several specific features. Its taxonomic spectrum is much more diverse than that in sample 30, two meters below. It contains *Cyclonephelium ordinatum, Cyclonephelium* spp.,

Cordosphaeridium, Impagidinium, Cleistrosphaeridium, Operculodinium plactum, Operculodinium f. A, *Hystrichokolpoma, Spiniferites ramosus, Spiniferites* spp., cf. *Dapsilidinium*, and *Rhombodinium*. The acritarchs are represented by *Micrhystridium*. But neither *Chiropteridium partispinatum* nor *Rhombodinium draco*, typical of next zonal assemblage, were found. This definitely implies that, starting at this level, a connection between the brackish-water basin and the open ocean originated.

Two features should be indicated to sum up the description of this zonal assemblage. (1) All samples recovered from the Solenaya Balka Horizon contain amorphic organic matter, and the greater is its share in the phytoplankton composition, the poorer is the assemblage. This may imply a higher or lower oxidation state of the basin and concerns samples recovered from the top of the Solenaya Balka Horizon, directly below the regressive sand sequence, and partly from the ostracod beds. (2) The persistent presence, despite their status as rare specimens, of dinocysts usually dwelling in open sea basins with normal salinity (particularly in samples 55, 50, 45, and 40) indicates that originated the connection between the Solenaya Balka basin and open ocean was permanent though troublesome.

Assemblage with Chiropteridium partispinatum. Samples 28, 27, 26, 20, 10, and 1 (depth interval of 130–48 m).

As soon as the normal salinity environment was restored, the abundance and diversity of phytoplankton grew. The distribution range of the assemblage is divided into two parts, the upper and the lower. In addition to zonal species, the lower part contains *Rhombodinium draco* as a main component, associated with *Gerdiocysta* and *Cyclonephelium* inherited from the brackish-water deposits (Member VI and the basal part of Member VII of the Kalmyk Formation, samples 28, 27 and 26; depth interval of 130–94 m). In the upper part (the larger part of Member VII, samples 20, 10, and 1; depths levels of 84, 64 and 50 m), main components are *Ch. partispinatum, Deflandrea spinulosa*, and less abundant *Homotryblium tenuispinosum* and *Pentadinium laticinctum.*

The entire Kalmyk Formation, and especially Member VII, is characterized by the presence of particular forms, redeposited from underlying deposits. This member, as mentioned before, is of the late Oligocene age (samples 20–10 and 1 with *D. spinosa, Ch. partispinatum, H. teuispinosum* and *P. laticinctum*). Member VI with *Ch. partispinatum* and *R. draco* may belong to both the upper part of the lower Oligocene and the upper Oligocene, as it is substantiated only by rare zonal species.

In addition to *R. draco,* lower beds with *Ch. partispinatum,* the assemblages contain rare specimens of *Homotryblium, Oligosphaeridium, Hystrichokolpoma, Cordosphaeridium, Membranophoridium aspinatum, Gerdiocysta, Cyclonephelium, Spiniferites*, and *Tetraporina.* Individual palinomorphs in the lower part of the section may belong to redeposited forms.

In samples 27 and 26, phytoplankton is more diverse and abundant (more than 50 taxons representing about a quarter of all palynomorphs). Among wetzeliellas, there are rare spicemens of *W. gotchii* and *W. symmetrica* and more than ten specimens of cf. *Wetzeliella articulata.* They coexist with *R. draco* (more than 20 specimens) and other species of the genus *Rhombodinium,* and also with *Cyclonephelium ordinatum, C. pastielsii* and *Cyclonephelium* sp. The abundance of *Lingulodinium, Operculodinium placitum, Impagidinium, Spiniferites*, and *Ch. partispinatum* corresponds to dozens of specimens, whereas single finds are characteristic of *Gardiocysta, Palaeocystodinium golzowense, Pentadinium laticinctum, Deflandrea phosphoritica, Hystrichokolpoma salacia, Dapsilidinium simplex, Cordosphaeridium exuberans, Areoligera, Kallosphaeridium, Cribroperidinium, Apteodinium* (including *A. emslandense*), and *Selenopemphix nephroides.* The group of prasinophytes and acritarchs includes *Crassosphaera, Tetraporina, Tytthodiscus, Cymatiosphaera, Ptanospermella, Micrhystridium, Ovoidites,* and *Pediastrum.* The same interval of the Kalmyk Formation was penetrated in Borehole 1, drilled north of Borehole 45A near the Tsatsa Lake. It is easily recognizable owing to presence of *Ch. partispinatum* and *R. draco* (samples 23, 24; depth interval of 60–70 m).

Member VII contains *Ch. partispinatum* and rather abundant *Deflandrea spinulosa* and *Homotryblium* from *H. floripes–H. tenuispinosum* group (particularly in sample 10). Accessory taxons are *Hystrichokolpoma, Lingulodinium, Chiropteridium, Homotryblium, Wetzeliella, Pentadinium laticinctum, Rhombodinium, Apteodinium maculatum, Membranophoridium aspinatum, Spiniferites, Systematophora, Thalassiphora pelagica, Deflandrea phosphoritica, Dapsilidinium, Tectatodinium, Selenopemphix nephroides, Operculodinium,* cf. *Muratodinium,* and *Gerlachodinium aechmophorum.* Single finds of acritachs and prasinophytes are classed with *Ptanospermella, Tetraporina, Pediastrum*, and *Cymatiosphaera.* Some of these components may be redeposited. Those taxons that are represented by rare specimens and disappear by the end of the Oligocene give grounds to state with sufficient confidence that Member VII should still be referred to the Oligocene. This inference is in agreement with the conclusion of Ben'yamovskii that upper horizons of Member VII of the Kalmyk Formation (depth interval of 86–48 m) correspond to the foraminiferal zone of *Cibicides ornatus.* Earlier, this zone was also identified in the North Caucasus sections (Zelenchuk Formation and the lower beds of Karadzhalga Formation) and was considered as a part of the Caucasus regional stage correlative with the lower Miocene. I do not argue against possible correlation between the upper part of Member VII of the Kalmyk Formation and the Caucasus regional stage, whose phytoplankton assemblage also contains a notable amount of *Deflandrea spinulosa,* but I consider it

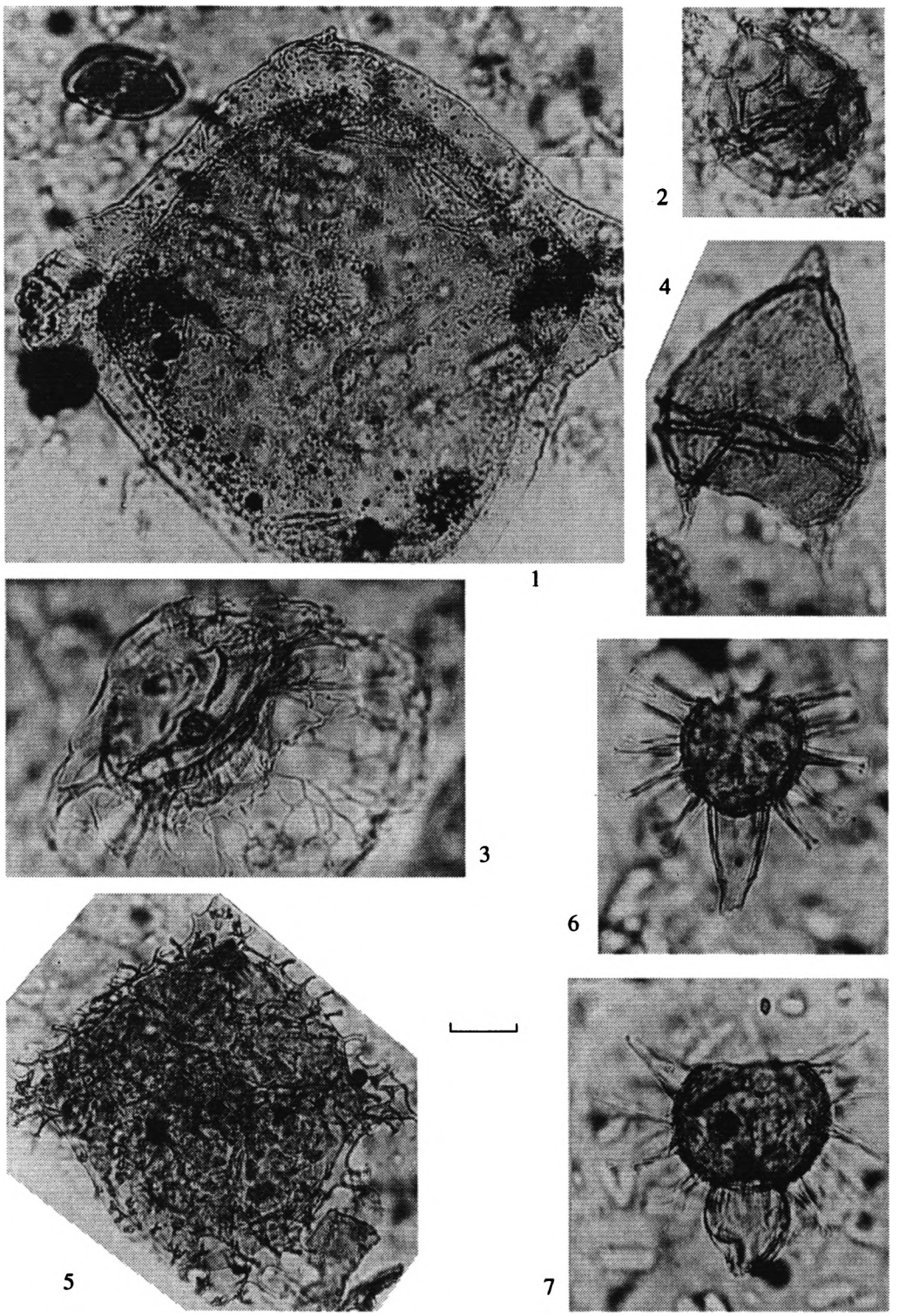

Plate I. Eocene–Oligocene phytoplankton from Borehole 45A: (1) *Rhombodinium perforatum* (Jan du Chene and Chateauneuf) Lentin and Williams, sample 110, Tsimlyansk Formation; (2) *Phthanoperidinium amoenum* (Drugg and Loeblich), sample 115, Tsimlyansk Formation; (3) *Adnatosphaeridium* sp., sample 110, Tsimlyansk Formation; (4) *Lentinia serrata* Bujak, sample 110, Tsimlyansk Formation; (5) *Wetzeliella symmetrica* Weiler, sample 65, Tsimlyansk Formation; (6) *Diphyes colligerum* (Dellandre and Cookson) Cookson, amend. Goodman and Witer, sample 144, Cherkessk Formation; (7) *D. ficusoides* Islam, sample 144, Cherkessk Formation; bars in all figures are equal to 15 μm.

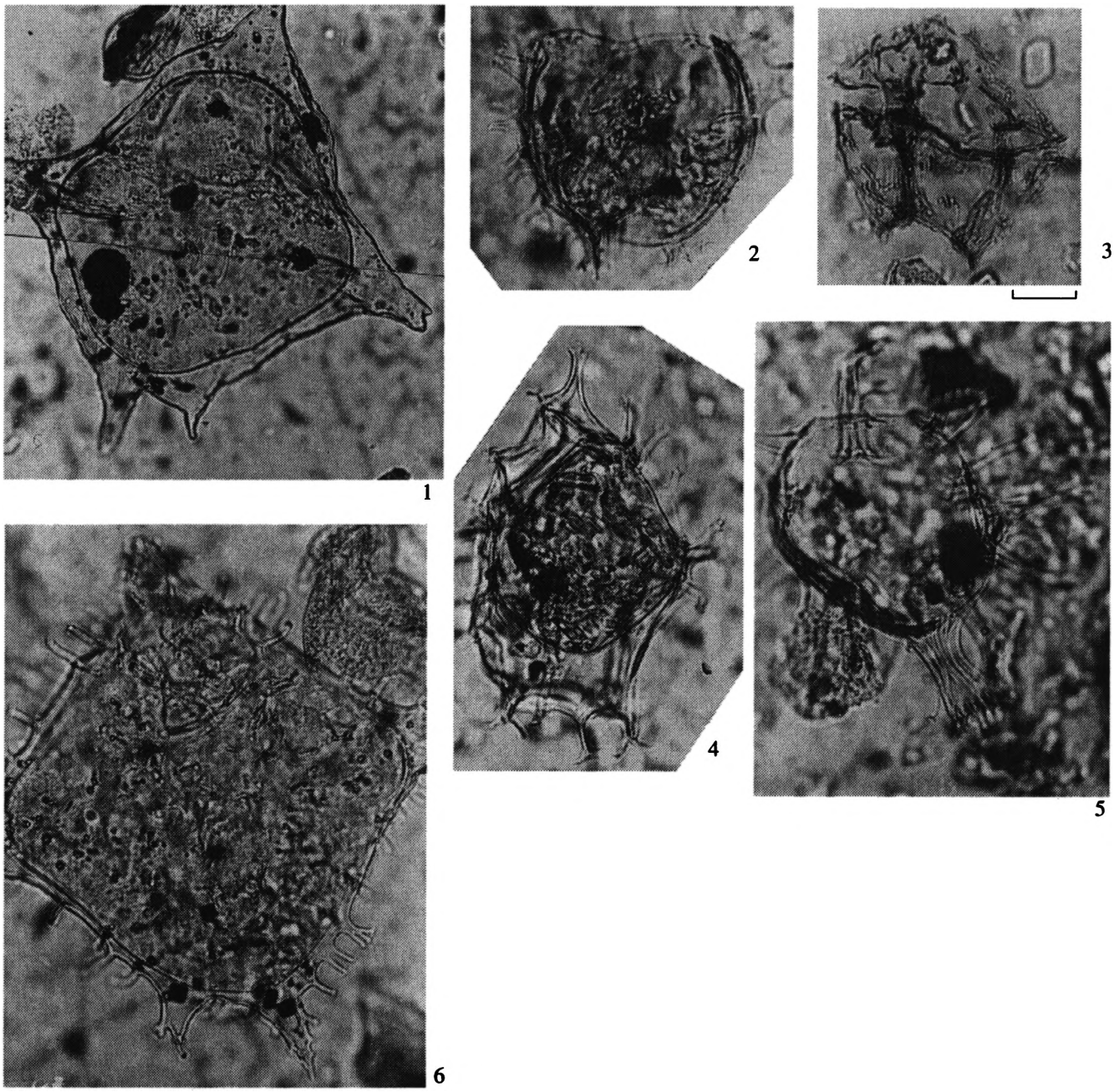

Plate II. Eocene–Oligocene phytoplankton from Borehole 45A: (1) *Rhombodinium draco* Gocht, sample 26, Kalmyk Formation; (2) *Chiropteridium partispinatum* (Gerlach) Brosius, sample 1, Kalmyk Formation; (3) *Gerlachidium aechmophorum* (Benedek) Benedek and Sarjeant, sample 10, Kalmyk Formation; (4) *Rottnestia borussica* (Eisenack) Cockson and Eisenack, sample 144, Cherkessk Formation; (5) *Homotryblium floripes* (Deflandre and Cookson) Stover, sample 80, Tsimlyansk Formation; (6) *Wetzeliella articulata* Eisenack, sample 26, Kalmyk Formation; bars in all figures are equal to 15 μm.

impossible to refer this regional stage, or at least its lower substage, to the lower Miocene.

The study of pollen and spores from the Oligocene section also revealed three palynological assemblages, as with phytoplankton; the lower and upper assemblages were recognized in marine deposits and the middle one in brackish-water sediments. According to local stratigraphic subdivisions, these palynological units were called the Tsimlyansk, Solenaya Balka, and Kalmyk assemlages.

In the Tsimlyansk assemblage, corresponding to basal brackish-water deposits with *Ergenica cimlanica,*

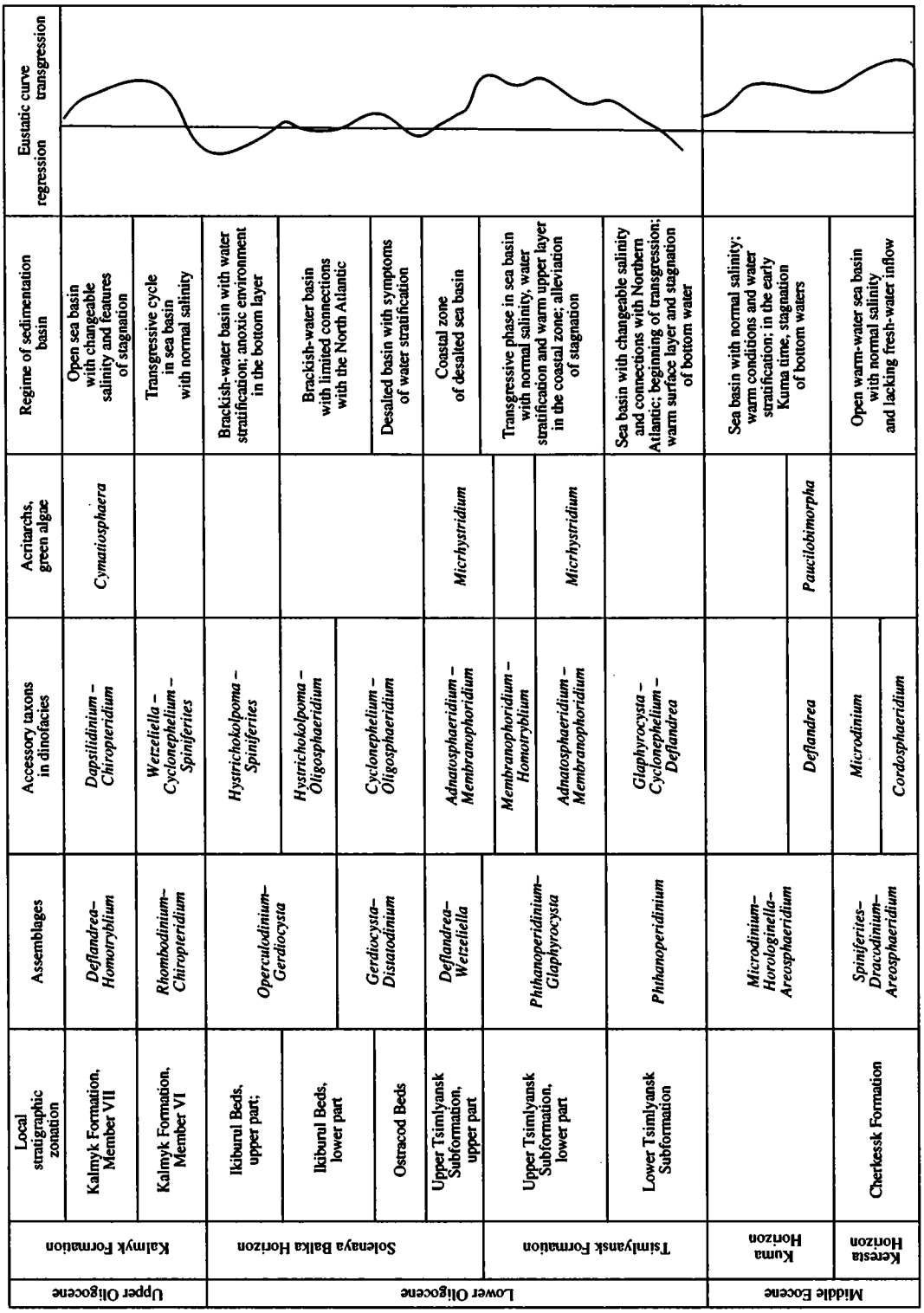
Upper Oligocene
Lower Oligocene
Middle Eocene
Kalmyk Formation
Solenaya Balka Horizon
Tsimlyansk Formation
Kuma Horizon
Keresta Horizon
Local stratigraphic zonation
Kalmyk Formation, Member VII
Kalmyk Formation, Member VI
Ikiburul Beds, upper part;
Ikiburul Beds, lower part
Ostracod Beds
Upper Tsimlyansk Subformation, upper part
Upper Tsimlyansk Subformation, lower part
Lower Tsimlyansk Subformation
Cherkessk Formation
Assemblages
Deflandrea–Homotryblium
Rhombodinium–Chiropteridium
Operculodinium–Gerdiocysta
Gerdiocysta–Distatodinium
Deflandrea–Wetzeliella
Phthanoperidinium–Glaphyrocysta
Phthanoperidinium
Microdinium–Horologinella–Areosphaeridium
Spiniferites–Dracodinium–Areosphaeridium
Accessory taxons in dinofacies
Dapsilidinium – Chiropteridium
Wetzeliella – Cyclonephelium – Spiniferites
Hystrichokolpoma – Spiniferites
Hystrichokolpoma – Oligosphaeridium
Cyclonephelium – Oligosphaeridium
Adnatosphaeridium – Membranophoridium
Membranophoridium – Homotryblium
Adnatosphaeridium – Membranophoridium
Glaphyrocysta – Cyclonephelium – Deflandrea
Deflandrea
Microdinium
Cordosphaeridium
Acritarchs, green algae
Cymatiosphaera
Micrhystridium
Micrhystridium
Paucilobimorpha
Regime of sedimentation basin
Open sea basin with changeable salinity and features of stagnation
Transgressive cycle in sea basin with normal salinity
Brackish-water basin with water stratification; anoxic environment in the bottom layer
Brackish-water basin with limited connections with the North Atlantic
Desalted basin with symptoms of water stratification
Coastal zone of desalted sea basin
Transgressive phase in sea basin with normal salinity, water stratification and warm upper layer in the coastal zone; alleviation of stagnation
Sea basin with changeable salinity and connections with Northern Atlantic; beginning of transgression; warm surface layer and stagnation of bottom water
Sea basin with normal salinity; warm conditions and water stratification; in the early Kuma time, stagnation of bottom waters
Open warm-water sea basin with normal salinity and lacking fresh-water inflow
Eustatic curve
regression
transgression

Fig. 5. Assamblages of dinocysts forming dinofacies in the section of Borehole 45A.

gymnosperms (usually 80–90%) are more abundant than angiosperms; Taxodiaceae prevail over Pinaceae by four to one in the lower part of the section and three to one or two to one in the upper part. Pollen of *Betula alba* dominates in the spectrum of Amentiflorae.

The Solenaya Balka assemblage is characteristic of the entire brackish-water sequence (196–132 m) and of the lower part of the Kalmyk Formation (down to 98 m). It is rich in gymnosperms (angiosperms make up no more than 57–75%) and shows the following features: a predominance of Pinaceae over Taxodiaceae (two to one, or three to two, except for sample 45, where Taxodiaceae are somewhat more numerous); a notable presence of *Sciadopytis* pollen; an increased significance of Fagaceae; and, in general, a larger amount of thermophilic components and a higher taxonomic diversity of pollen. This part of the section was probably accumulated in a period of climatic warming, when the mean annual precipitation rate was lower than before.

The Kalmyk assemblage from the upper marine part of the Oligocene section is more rich in gymnosperms (up to 85%) and shows the alternating prevalence of either Pinaceae or Taxodiaceae and the dominant position of *Betula* among Amentiflorae. This may imply the beginning of a cooling stage in the late Oligocene.

In Tsimlyansk assemblage, as mentioned earlier, the Taxodiaceae dominate over Pinaceae in Member II and, to a lesser degree, in Member III; Betulaceae, particularly *Carpinus,* also belong here to dominant taxons. The assemblage is composed of typical lower Oligocene taxons of the southern Russian platform, Usturt, Aral area, Turgai region, and southern Siberia. These are *Picea tobolica, Tsuga crispa, Ephedra cheganica, Betula gracilis, B. trigona, Juglans polyporata, J. sibirica, J. sieboldianiformis, Tilia tomentosiformis, Carya glabraeformis,* and *C. spackmania.* Less abundant *Rhus, Comptonia,* and, as usual, rare specimens of more thermophilic Moraceae, *Ilex, Nyssa, Liquidambar, Carylopsis, Hamamelis,* and other taxons are also present. The content of spores does not exceed 1–2% (*Cyathea, Lygodium, Osmunda,* Polypodiaceae, and *Lycopodium*).

In the Solenaya Balka assemblage, where Pinaceae dominate over Taxodiaceae, there is a notable amount of *Sciadopytes,* and Fagaceae prevail here among Amentiflorae (25–40% of the assemblage composition). The abundance of thermophilic components is significant, and their forms listed in the lower assemblage coexist here with *Alfaroa, Trochodendron, Engelhardtia, Cyclocarya,* Sterculiaceae, *Palmae, Aralia,* and *Castanopsis.* There are many pollen species typical of the lower Oligocene and inherited from the lower assemblage; among them are *Pinus koraensis, P. aralica, Tsuga turulosa, T. crispa, T. canadensiformis, Ephedra cheganica, Betula trigona, B. gracilis, Alnus* cf. *minor, Corylus granolata, Juglans polyporata, J. siboldianiformis, J. sibirica, Carya glabraiformis, Quercus graciliformis, Q. gracilis, Q. conferta, Q. sibirica, Q. williamsonia, Q. forestdalensis* (it is remarkable that the fine pollen of oaks of evidently evergreen species makes up the main part in the Fagaceae group, especially in samples from ostracod beds), *Nyssa crassa, N. rotundata, Rhus ustjurtensis, Liquidambar zaissanica, Tilia tomentosiformis, Fothergilla gracilis, Corylopsis crassa,* and others. The abundance of broad-leaved plants in samples is different: Sample 60 is rather rich (>5%) in *Ulmus, Rhus,* and *Tilia;* sample 55 contains *Liquidambar;* sample 45 includes *Rhus* and *Liquidambar;* sample 40 bears *Myrica;* and samples 30 and 29 yielded *Comptonia, Myrica, Nyssa,* Moraceae, Ericaceae and *Ilex.* The upper interval of brackish-water Member V is rich in spores of ferns, particularly of *Osmunda.* Apparently, the upper 20 meters of this interval corresponds to a new stage of higher humidity. Ferns, Myricaceae, Ericaceae, and Nyssaceae apparently were dominant in coastal vegetation that populated the dried up areas of former sea floor after regression.

The Kalmyk assemblage (samples 20, 10, 1) is also rich in gymnosperms (80–85%) alternately dominated by either Taxodiaceae or Pinaceae; Betulaceae are also abundant, particularly *Alnus,* but thermophilic components with typical Oligocene taxons are less widespread, mainly among Amentiflorae. The abundance of fine oak, especially of their evergreen representatives is reduced, while the role of *Fagus* is more significant. The amount of fern spores (*Osmunda* and others) is again reduced. The assemblage is quite typical of the upper Oligocene deposits, but not of their uppermost horizons, which already lack the characteristic Juglandaceae species.

CONCLUSION

Phytoplankton and palynomorphs from Borehole 45A are used to more accurately date separate intervals of the section and to correlate them with coeval units in neighbouring regions. The Tsimlyansk Formation correlates with the Pshekha Horizon of the North Caucasus sections, excluding its basal layers, and with the Ashcheairyk Formation of northern Usturt and the Borysthenes Horizon of the Black Sea region. The phytoplankton assemblage from ostracod beds, though less diverse than that of the Polba Horizon in the Kuban' section, is similar to it in general composition. Phytoplankton from the Ikiburul Beds includes a great variety of taxons, which are identical to those of the Serogozy Formation in the Black Sea region (Andreeva-Grigorovich, 1991) and to microflora of the Tamda Formation in northern Usturt (Popov *et al.,* 1993). Marine phytoplankton of the Kalmyk Formation is common in the Chattian deposits of Northern Germany, the Askania and Gornostaevka formations of the Black Sea region, and the Karatomak and Baigubek horizons of the Aral and Usturt areas. In the latter, the assemblage with *Chiropteridium partispinatum, Deflandrea spinulosa, Rhombodinium draco,* and *Homotryblium*

floripes was recovered from Borehole 3006 (Zaporozhets, 1991). The climate of the Solenaya Balka time was warm, seasonal and, on the whole, more arid, as can be judged from notable presence of Fagaceae (including *Castanopsis*), Palmae, Moraceae, Araliaceae, and various thermophilic deciduous and evergreen broad-leaved species in this polynoflora, whose background taxons belong to Pinaceae. In the middle Eocene (the Keresta or late Cherkessk time), the climate was subtropical, but cooler and more humid than in the Kuma time, when it was more seasonal with a hot and dry summer.

The analysis of phytoplankton assemblages suggests a conclusion concerning the hydrological regime of the past sea basin. The Tsimlyansk time corresponds to a period of normal salinity in the basin with well stratified water masses and anoxic environments near the bottom, especially in the early Tsimlyansk time. In the late Tsimlyansk time of transgression development, the surface water layer experienced desalination, and the basin became more shallow owing to the greater rate of sedimentation. This is indicated by the abundance of *Micrhystridium* species, sometimes constituting more than 50% of phytoplankton spectra (Fig. 5).

In the Solenaya Balka time, the basin was semisaline and partly isolated from the ocean. The present Black Sea can be considered as a model of that basin. The degree of its isolation and salinity was inconstant. The salinity increased in the period of better connection with the Atlantic Ocean, as is evident from the similar composition of molluscan and dinocyst assemblages in the Solenaya Balka Horizon, the Tard Clay of Hungary, and the Ileanda Beds in Transylvania basin of Romania and Czechoslovakia (Baldi, 1986; Rusu, 1988). *Ergenica cimlanica*, a typical mollusk from the basal beds of the Solenaya Balka Horizon, was also found in different regions of Central Europe (near Budapest in the Transylvania basin, and also in Slovenia and Bulgaria). The basin in question could have been connected with the Atlantic Ocean through the southern termination of the Mainz depression, as suggested by Rusu (1988). The anoxic environment in the Solenaya Balka basin is inferred from a high content of amorphic organic matter in the studied preparations (up to 80–90% of the total amount). This feature is particularly typical of the ostracod beds and the upper Ikiburul strata. In the early and middle Ikiburul time, the connection with ocean was more distinct. The Solenaya Balka basin was desalted not only due to the limited discharge of fresh waters, but also as a result of general desalting of surface waters of the World Ocean, as suggested by Muzylev *et al.* (1992). As testified by the composition of phytoplankton assemblages, the basin of the Kalmyk time had normal salinity and a permanent connection with the Atlantic; apparently, its waters were also distinctly stratified.

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Subdivision of Oligocene and Lower Miocene Sediments of Northern Sakhalin Evidenced by Mollusks

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Abstract—Upper Eocene, Oligocene, and lower Miocene units were distinguished in the Lower Cenozoic Machigar and Tumi formations of the Schmidt Peninsula, northern Sakhalin, using distribution patterns of 77 mollusk species. A detailed study of their assemblages permitted for the first time the subdivision of the section into nine faunal beds. The results are used to correlate the section with coeval deposits in adjacent regions and to perform paleogeographical reconstructions.

Key words: *northern Sakhalin, Oligocene, lower Miocene, mollusk beds, paleogeography, paleoclimate.*

INTRODUCTION

A series of problems concerning Oligocene deposits in the North Pacific still remain unsolved. Paleontological remains from the Oligocene sequences of different regions are not thoroughly studied, the boundaries of Oligocene stages are not clearly determined, and paleogeographic environments and geologic events of that time are poorly understood. That is the reason why the detailed study of Far East sections is nowadays of particular importance. Sakhalin and Kamchatka sections, characteristic of the continent–ocean transition zone, are among those attracting considerable attention. In northern Sakhalin, one section of the Schmidt Peninsula is one of the most representative in that series (Fig. 1). It is complete and rich in paleontological remains.

The section includes pre-Cenozoic, namely, Jurassic and Cretaceous, and Cenozoic deposits, the latter spanning an interval from the upper Eocene–Oligocene to the Pliocene. The Cenozoic was studied here at different times by N.N. Tikhonovich, P.I. Polevoi, E.M. Smekhov, I.I. Ratnovskii, N.B. Vassoevich, G.S. Mishakov, V.I. Grechin (Ratnovskii, 1960; Vassoevich, 1961; Grechin, 1976), and others. Biostratigraphic investigations were made by Khomenko (1934, 1938), Voloshinova *et al.* (1970), Krishtofovich (1964), Gladenkov (1973), and Zhidkova *et al.* (1981). Later, researchers from the Geological Institute, Russian Academy of Sciences, obtained micropaleontological characteristics of the sediments (Oreshkina and Vitukhin, 1987).

Below, we present data on the Oligocene–lower Miocene part of the section. This part, described by Gladenkov on the northern coast of the Schmidt Peninsula, has its lower contact located 4.5 km southeast of the Mariya Cape and its upper one detected eastward, near the Monchigar Lake. For comparison, we use data on coeval sequences from the western coast of the peninsula, northwest of the Pil'vo River mouth. In the paleontological characterization of the studied sediments, we focused our attention on mollusks that are considerably diverse and represented by more than 70 recorded species.

Cenozoic mollusks from the Schmidt Peninsula were studied at different times by I.P. Khomenko,

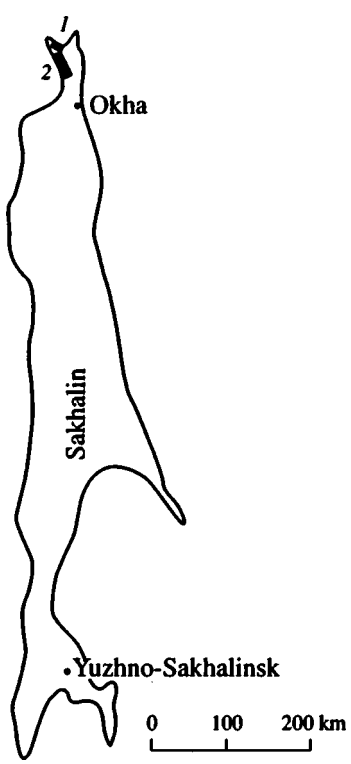


Fig. 1. Studied sections in the northern Sakhalin: (*1*) Machigar section; (2) Tumi section.

F.G. Lautenshleger, N.Ya. Kogan, I.N. Kuzina, L.V. Krishtofovich, L.A. Popova, Yu.B. Gladenkov, L.S. Zhidkova, and others. In this paper we use the materials of collections gathered by Gladenkov in 1969 and 1979. The bed-by-bed study of molluscan assemblages enabled a more detailed subdivision of the section and provided a new basis for its correlation.

DESCRIPTION OF THE SECTION

Paleogene and lower Neogene sediments of the Schmidt Peninsula are studied in two reference sections: (1) the Machigar section on the marine coast of the Severnyi Bay northwest of the Monchigar Lake; and (2) the Pil' section on the marine coast of the Sakhalin Bay near the Vodopadnaya River mouth. The former includes the basal Machigar (sequences I–VI) and Tumi (sequences VII–XI) formations. A description of the formations and their sequences in the stratigraphic succession is given below. Within the sequences we distinguish members numbered separately for each formation.

MARINE COAST IN THE MONCHIGAR LAKE REGION (MACHIGAR SECTION)

The Machigar Formation was distinguished in 1935 by N.S. Erofeev. Its lower contact is of unclear type. The formation seems to be separated from the Cretaceous volcanic rocks by a tectonic contact, and based on the geologic position, it overlies these rocks unconformably with a hiatus (Fig. 2).

Sequence I (Coal-bearing, Sandy)

1. Sandstone, sand, and clay beds. The sandstone beds are greenish gray, medium-grained, loose, sometimes clayey, and contaminated with dispersed gravel and hard concretionary layers. The horizons contain siderite concretions from 0.3–0.5 to 1 × 5 m in size along with lenses and interlayers of brown coal up to 2–2.5 m thick. The total thickness is up to 15–20 m. According to A.I. Poyarkova, floral remains of the unit are represented by *Osmunda sachalinensis, Taxodium dibium, Betula* cf. *elliptica, Fagus* sp., and *Magnolia* sp. (cited in Ratnovskii, 1960).

2. Yellow gray and red sandstones, coarse- to medium-grained, thin- and cross-bedded, tuffaceous, bearing coaly pebbles, plant detritus, coal lenses 0.2–0.3 m thick, clayey interbeds, and jarosite. The lower part contains carbonate concretions 0.2–0.5 × 1–1.5 m in size. The member yielded remains of the following mollusks: *Mytilus littoralis, Corbicula sitakaraensis,* and *Cerithidea quadrimonilicosta.* The thickness is 10–12 m.

Sequence II (Aleurolitic-sandy)

3. Gray fine-pebble conglomerates and blue-green loose and hard gravelstones 2–4 m thick occur at the base of the member. Subrounded pebbles are mainly composed of extrusive rocks. Conglomerates contain shells of *Papyridea harrimani.* These rocks grade upward into dark gray, brown siltstones, nonbedded, fine-rubble, rather soft, displaying a shelly parting and sand admixture. A layer of carbonate concretions 0.5 × 3 m in size, bearing dispersed pebble and plant detritus, is located at the base. The sediments contain remains of *Macoma sejugata, Musculus* cf. *snatolensis, Acanthocardia* cf. *snatolensis, Portlandella watasei,* and other mollusks. The thickness is up to 20 m.

Members 1–3 are exposed in flanks of a small anticline and in a synclinal fold in three sections. The following members are exposed in a monocline dipping northeastward.

4. Sandstones with interbeds of silty conglomerate (0.7 m thick) and coal lenses predominate in the lower part, whereas the upper one is composed of greenish sandstone, medium-grained, silty, ferruginate, slightly bedded, bearing abundant pebbles and dispersed concretions 1 × 1.5 m in size. The first gennoishi (glendonite) inclusions occur at the top. The thickness is 30 m.

Sequence III (Conglomerate-Sandy, with 16 Cycles)

5. Basal horizon (2 m thick) is represented by gray conglomeratic sandstone bearing large carbonate concretions (up to 2 × 4 m in size) and remains of the following mollusks: *Nemocardium iwakiense, Thracia schmidti, Papyridea harrimani, Mya grewingki,* and others. It is the first marker horizon with abundant fauna. Higher beds (5.5 m thick) are represented by blue-green conglomeratic sandstone bearing abundant jasperoid pebbles and concretions 0.5–0.7 × 3 m in size; the rocks grades upward into pebbly sandstones and contaminated siltstones with small (0.2 m) concretions and gennoishi. The overlying part consists of alternating conglomerate (up to 1 m), sandstone, and siltstone (up to 3–3.5 m) beds constituting cycles of about 5 m thick. Concretions of various sizes, from 0.1 to 1 m, gennoishi, pebble, and small boulders of different composition, among them granitoids, are present at this level. The lower contacts of conglomerate beds are commonly uneven and erosional, though some of them are gradual. This part of the section yielded abundant mollusks *Liocyma furtiva, Papyridea matschigarica, Cyclocardia matschigarica, Trachycardium kinsimarae, Ciliatocardium matchgarense, Arca sakamizuensis, Pododesmus schmidti, Modiolus matchgarensis, Periploma besshoensis,* and others. Abundant remains of large-valved mollusks are common in conglomerates. Different species of particular genera, namely, *Ciliatocardium, Papyridea,* or *Mya,* are associated with certain beds. Mollusk accumulations are missing in siltstones; the latter contain isolated thin-walled forms

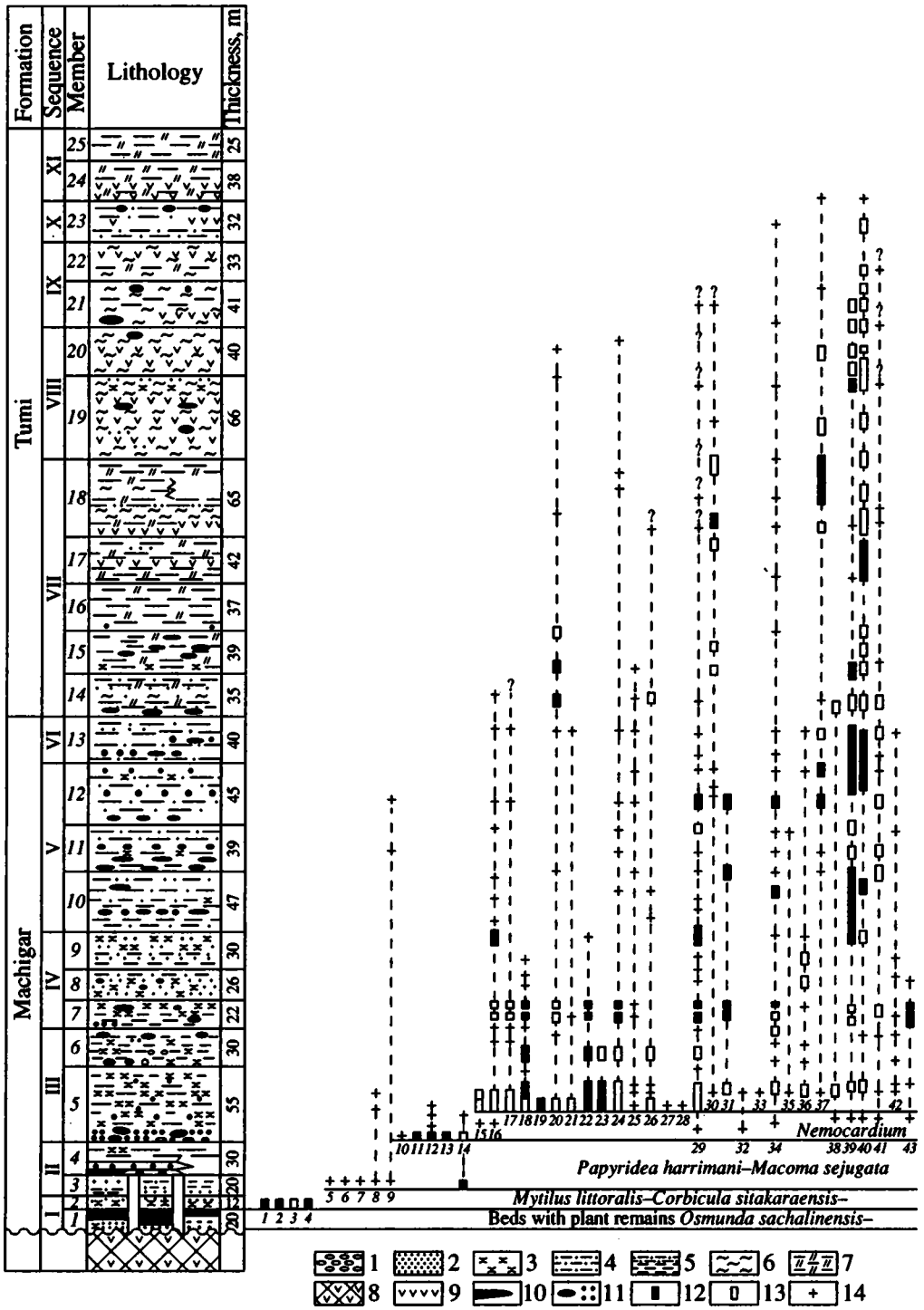
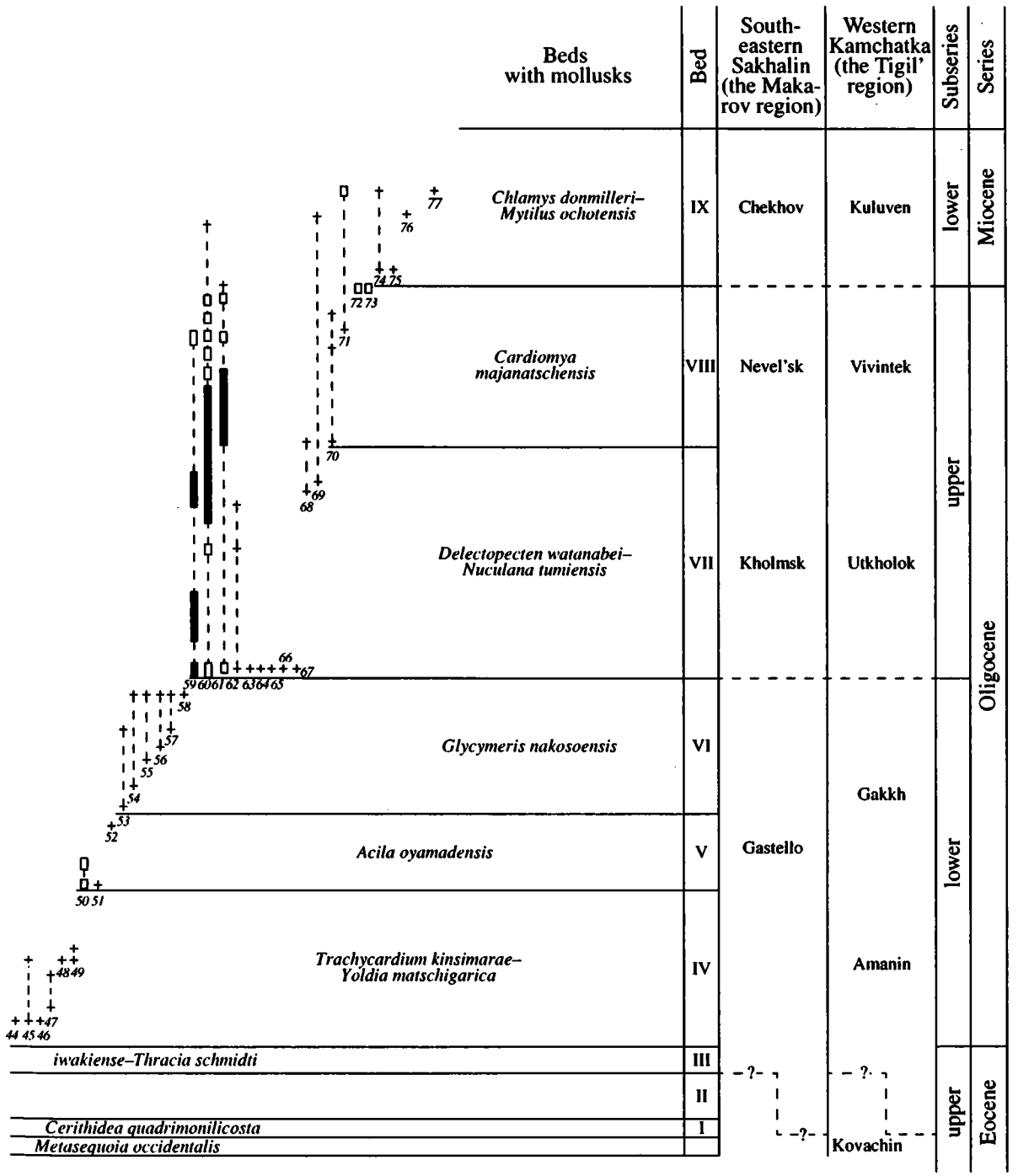


Fig. 2. Distribution of mollusks in the Machigar section: (1) conglomerate; (2) sandstone; (3) massive sandstone; (4) siltstone; (5) clay; (6) opoka; (7) tuffaceous siltstone; (8) volcanic rock; (9) tuff; (10) coal; (11) concretions; (12–14) number of molluscan shell remains: (12) over 10 specimens; (13) over 5 specimens; (14) from 1 to 5 specimens.

Mollusks: (*1*) *Mytilus littoralis* Slodkewitsch; (*2*) *Corbicula sitakaraensis* Suzuki; (*3*) *Corbicula* sp.; (*4*) *Cerithidea quadrimonilicosta* Khomenko; (*5*) *Acanthocardia* cf. *snatolensis* L. Krishtofovich; (*6*) *Musculus* cf. *snatolensis* Sinelnikova; (*7*) *Clinocardium* sp.; (*8*) *Portlandella watasei* (Kanehara); (*9*) *Macoma sejugata* (Yokoyama); (*10*) *Clinocardium taracaicum* (Yokoyama); (*11*) *Nemocardium iwakiense* (Makiyama); (*12*) *Mya grewingki* Makiyama; (*13*) *Thracia schmidti* L. Krishtofovich; (*14*) *Papyridea harrimani* Dall; (*15*) *Liocyma furtiva* (Yokoyama); (*16*) *Yoldia caudata* Khomenko; (*17*) *Cyclocardia matschigarica* Khomenko; (*18*) *Papyridea matschigarica* Khomenko; (*19*) *Arca sakamizuensis* Hatai and Nisiyama; (*20*) *Clinocardium makiyamae* Kamada; (*21*) *Ancistrolepis modestoideus* (Takeda); (*22*) *Yoldia matschigarica* L. Krishtofovich; (*23*) *Trachycardium kinsimarae* (Makiyama); (*24*) *Pododesmus schmidti* L. Krishtofovich; (*25*) *Turritella importuna* Yokoyama; (*26*) *Neverita asagaiensis* (Makiyama); (*27*) *Colus asagaiensis* Makiyama; (*28*) *Buccinum* sp.; (*29*) *Ciliatocardium matchgarense* (Makiyama); (*30*) *Yoldia laudabilis* Yokoyama; (*31*) *Ciliatocardium asagaiense* (Makiyama); (*32*) *Nuculana* sp.; (*33*) *Nucula matschigarica* L. Krishtofovich; (*34*) *Modiolus matchgarensis*



(Makiyama); (*35*) *Ciliatocardium yamasaki* (Makiyama); (*36*) *Mya cuneiformis* (Böhm); (*37*) *Conchocele smekhovi* (Kogan); (*38*) *Macoma asagaiensis* Makiyama; (*39*) *Periploma besshoensis* (Yokoyama); (*40*) *Macoma simizuensis* L. Krishtofovich; (*41*) *Portlandella nitida* (Slodkewitsch); (*42*) *Macrocallista* sp.; (*43*) *Chlamys matchgarensis* Makiyama; (*44*) *Lithophaga tumiensis* (Lautenschläger); (*45*) *Malletia* sp.; (*46*) *Ainicardita napanica* Popov; (*47*) *Ostrea* sp.; (*48*) *Ciliatocardium schmidti* (Khramova); (*49*) *Crepidula* cf. *matajiroi* Makiyama; (*50*) *Acila oyamadensis* Hirayama; (*51*) *Nuttalia* cf. *uchigoensis* Kamada; (*52*) *Keenocardium* sp.; (*53*) *Glycymeris nakosoensis* Hatai and Nisiyama; (*54*) *Musculus* sp.; (*55*) *Cyclocardia pilvoensis* Khomenko; (*56*) *Peronidea* cf. *ochii* Kamada; (*57*) *Yoldia* sp.; (*58*) *Dentalium* sp.; (*59*) *Delectopecten watanabei* (Yokoyama); (*60*) *Nuculana tumiensis* (Lautenschläger); (*61*) *Nucula tumiensis* L. Krishtofovich; (*62*) *Nuculana praerobai* L. Krishtofovich; (*63*) *Plicifusus* sp.; (*64*) *Fulgoraria* sp.; (*65*) *Crepidula* sp.; (*66*) *Cerithidea* sp.; (*67*) *Macoma* sp.; (*68*) *Macoma optiva* Yokoyama; (*69*) *Glycymeris wishkahensis* Clark; (*70*) *Cardiomya majanathensis* (Ilyina); (*71*) *Chlamys* ex gr. *cosibensis* Yokoyama; (*72*) *Chlamys donmilleri* MacNeil; (*73*) *Chlamys rikinensis* Sinelnikova; (*74*) *Cyclocardia* sp.; (*75*) *Acila* sp.; (*76*) *Lucinoma acutilineata* (Conrad); (*77*) *Mytilus ochotensis* Slodkewitsch.

(paleotaxodonts) buried in concretions. The total thickness is up to 55 m.

6. A similar alternation of conglomerate (0.5–1.5 m), sandstone, and siltstone (3–5 m) beds in cycles of about 5–6 m in thickness. Gennoishi are missing here, but round loaflike carbonate concretions up to 1.5 × 3.5 m in size occur, together with spheroidal, small, up to egg-sized, concretions and abundant pebbles. Numerous shells of *Yoldia matschigarica, Papyridea matschigarica, Ciliatocardium matchgarensis, Neverita asagaiensis, Pododesmus schmidti,* and others occur in the member. The thickness is up to 29–30 m.

Sequence IV (Sandy-Aleurolitic, with 10–12 Cycles)

7. Conglomerate, sandstone, and siltstone beds with abundant mollusks. The basal conglomerate bed (0.7–1 m thick) with the lower uneven boundary is medium-pebble, hard, bearing large carbonate concretions (up to 1 × 5 m) and abundant mollusks *Modiolus matchgarensis, Papyridea matschigarica, Chlamys matchgarensis, Yoldia matschigarica, Pododesmus schmidti, Ciliatocardium asagaiense,* and *C. matchgarense,* in association with remains of large balannuses and shell detritus. It is the second horizon with the accumulation of fauna. The overlying siltstone beds are 14 m thick and include a sandstone interbed (0.4 m) in the middle part, as well as carbonate concretions (up to 0.1 m), gennoishi, and mollusk remains. This bed grades into gray loose sandstone (3 m thick) with three denser interbeds bearing pebble and numerous remains of *Papyridea, Ciliatocardium, Chlamys,* and other mollusks. It is the third horizon with abundant fauna. The member is terminated by a siltstone bed 4 m thick. The total thickness is 22 m.

8. Alternation of sandstone (1.5–2 m) and siltstone (3–4–7 m) beds. The sandstone beds are greenish gray, medium-grained, sometimes bearing pebbles and detritus. The siltstone beds are ferruginate and nonbedded. Carbonate concretions in the beds are large, lenslike, up to 1 × 4 m in size or smaller (0.1–0.3 m), and spheroidal. The molluscan fauna is represented by *Ciliatocardium matchgarense, Mya cuneiformis, Chlamys matchgarensis, Papyridea matschigarica,* and others. The thickness is 25–26 m.

9. Interlayering of the similar sandstone (1–2 m), sometimes conglomeratic, and siltstone (3.5–7 m) beds. Carbonate concretions varying in size from 0.1–0.4 to 0.5 × 1 m, and gennoishi (0.15 m) occur. The member yields abundant *Yoldia caudata, Periploma besshoensis,* and *Ciliatocardium matchgarense.* The thickness is 29–30 m.

Sequence V (Aleurolitic, with 20 Concretionary Horizons)

10. Siltstone beds (4.5–8 m thick) with horizons of carbonate spheroidal and loaflike concretions varying in dimensions from 0.1–0.2 to 0.4–0.5–0.7 m. The concretions sometimes include pebbles. Some concretions are as large as 2.5 × 3 m in size. Gennoishi commonly occurring within the concretions are characteristic of the member. Numerous *Macoma semizuensis, Periploma besshoensis,* and *Yoldia caudata, Modiolus matchgarensis* were detected in the bed. The thickness is 46–47 m.

11. Siltstone beds (3–10 m) with lenslike horizons of concretions 0.5–1 × 1–3 m in size. The deposits also contain small and large (up to 0.4–0.7 m) individual concretions and gennoishi, as well as pebbles. The concretions include granite clasts. Accumulations of *Periploma besshoensis* and *Portlandella nitida* were found. The thickness is 39 m.

12. Siltstone beds (4–9 m) bearing spheroidal and loaflike concretions (from 0.1–0.5 to 0.7 m), gennoishi, carbonate lenses (0.1–0.4 × 0.4-2.5 m) and small carbonate segregations. The deposits include pebbles, granite detritus, and numerous remains of *Macoma simizuensis, Periploma besshoensis,* and *Conchocele smekhovi.* The thickness is 44–45 m.

Sequence VI (Siltstone)

13. Siltstone beds (2–5 m) differing from the underlying strata due to the occurrence of bedded hard interlayers (2 m thick) with whitish rubble. The sediments enclose concretions varying in size from 0.1 to 0.5–0.7 m, abudant gennoishi, and a carbonate bed 0.2–0.3 m thick at the top of the member. Accumulations of *Macoma simizuensis, Periploma besshoensis,* and *Conchocele smekhovi* are present. The thickness is 39–40 m.

The total thickness of the Machigar Formation is 415 m.

Tumi Formation

The formation was distinguished as a part of the section by Ratnovskii in 1952. The name was first suggested by Smekhov in 1936. The lower boundary of the unit is conformable.

Sequence VII (Opoka-Aleurolitic, Bedded)

14. Brown bedded member. Large carbonate concretions (1 × 2 m), hard and solid, are concentrated at the base. The overlying rocks are represented by interlayering soft clayey siltstones (1–5 cm), light gray dense opoka beds, and tuffaceous siltstones (0.1–0.3 m) with coarse splinter parting. Opokas are porcelaneous if weathered; they enclose plant detritus, sand nests, basal conglomerate lenses (0.2 m), pebbles, and granitoid blocks (0.7 × 1.5 m), frequently enveloped in carbonates. The rocks also include shells, separate concretions (0.2 × 0.7 m), and gennoishi (up to 0.6–0.3–0.4 m). The molluscan assemblage consists of abundant *Delectopecten watanabei, Nuculana tumiensis, Periploma besshoensis, Neverita asagaiensis,* and others. Shark teeth were found as well. The thickness is 33–35 m.

15. Carbonate sandstone bed (0.7 m) with pebbles at the base grades into alternating hard tuffaceous siltstone (0.1–0.2 m) and jarosite clay (1–5 cm), probably replacing the volcanic ash. The member is characterized by three horizons of very hard and large carbonate concretions (0.52–1.0 × 1–3 m), separated by intervals of 3, 5, and 10 m. Concretions contain pebbles, plant detritus, large gennoishi (0.3 m), fauna of *Delectopecten watanabei, Periploma besshoensis, Macoma simizuensis,* and other mollusks. Worm tracks occur in the sedimentary matrix. The thickness is 34–39 m.

16. A similar bed interlayering but without giant concretions. Small concretions (up to 0.3–0.5 m) sometimes bear granite pebbles. The rocks are thin- and microbedded, sometimes sandy, bearing pebble and jarosite inclusions, and are tuffaceous, whitish, and porcelaneous when dried. The member contains abundant pebbles, fine plant detritus, hard interbeds of massive rocks (up to 0.3–0.52 m), and lenses with rare concretions (0.4 × 2 m) of cream-colored carbonates. The thickness is 37 m.

17. The similar rocks enclosing bluish, slightly bedded, dense, tuffaceous interbeds with coarse parting, and vitro-lithoclastic hard and loose (0.1 m) tuffs with occasional pumice. The unit demonstrates horizons of concretions (0.3–0.7–1 × 2 m), granite and basalt blocks, and interbeds of gennoishi aggregates (0.2 m). The rocks are somewhat gypsinate and bear abundant *Macoma simizuensis* and *Nuculana tumiensis*. The thickness is 42 m.

18. Alternating bluish tuffaceous siltstone and opoka beds with thin clayey interlayers. The member contains horizons of carbonate concretions (from 0.3–0.6 to 1–1.5 m), thin lenses of fine tuffaceous material, abundant jarosite, sandy interbeds and lenses, gennoishi, granite and basalt boulders, often enveloped in carbonates, and pebbles. The unit includes accumulations of *Nuculana tumiensis, Macoma simizuensis, Conchocele smekhovi,* and *Delectopecten watanabei.* The thickness is 60–65 m.

Sequence VIII (Opoka Beds)

19. In the Up'ef River mouth opokas are light gray to bluish, commonly massive, occasionally bedded, hard, fine-clastic, homogeneous, enclosing tuff (0.1 m) and clay interbeds after tuffs (5 cm thick and separated by intervals of 0.5 m). The sediments contain granite blocks (1.5 m) covered with carbonate, gennoishi, carbonate concretions (0.5 m in general and up to 2 × 4 m in the middle part), and a giant concretion 4 × 5 m in size consisting of 9–10 spheres. Concretion nuclei are often represented by volcanic clasts (up to 0.5 m) or molluscan shells. Abundant thin-walled mollusks *Nuculana tumiensis, Nucula tumiensis,* and *Macoma simizuensis* were found here. The thickness is 66 m.

20. Similar opoka beds bearing loose psammitic tuff and moistened clay interlayers (up to 0.1 m thick and separated by intervals of 0.5–1.5 m), gennoishi, spheroidal concretions (0.2–0.5 m), sand nests, dispersed pebbles, and granite clasts. The rocks are yellowish. In the basal part of the member, tuffs include gypsum crystals up to 5–7 cm in size. Its upper part encloses abundant jarosite and a carbonate lens (1 × 2.5 m). The sediments contain blocks of exotic rocks enveloped in carbonates (up to 4 m) with mollusks *Pododesmus schmidti, Modiolus matchgarensis,* and others. Abundant *Macoma simizuensis, Periploma besshoensis, Nuculana tumiensis, Conchocele smekhovi,* and fish scales were found in the unit. The thickness is 40 m.

Sequence IX (Whitish Opoka Beds)

21. Whitish siliceous hard opokas with shelly jointing, homogeneous, and porcelaneous. The rocks enclose tuff (0.1 m) and clay separated by intervals of 0.5–1.5 m beds, rare concretionary interlayers (0.2–0.3 m), and a carbonate lens (1 × 5 m) at the base. Plant detritus, pebbles, sandy varieties, and abundant jarosite occur. Rusty yellow patches of ferrugination and Liesegang rings were recorded at weathered rock fragments. The sediments contain numerous *Nuculana tumiensis, Periploma besshoensis,* and *Macoma simizuensis. Chlamys donmilleri, C. rikinensis,* and others are less abundant. The member locally shows a "zebra bedding," i.e., an alternation of light and yellowish brown rocks. The thickness is 41 m.

22. Gray opokas, dense, bedded, often porcelaneous, siliceous, bearing interbeds of clayey rocks and clays (0.2–0.3 m) separated by intervals of 0.3–1 m. In the lower and upper parts the member, there are interbeds of loose tuff (0.1–0.3 m) substituted by montmorillonite clay. Jarosite is common. A considerable part of the member consists of dark laminated and light, sometimes ferruginated opokas alternating with intervals of 0.2–0.3 m. The thickness is 33 m.

Sequence X (Siltstone)

23. Grey and bluish siltstone beds, soft, superficially slightly bedded, but thin-bedded inside. Tuff interbeds (0.1–0.2 m) occur in the lower and middle parts of the member. The sediments contain lenses of psammitic and psephitic tuff, rare pebbles, plant detritus, fish scales, fucoids, and hard yellow, probably silicified, concretions (up to 0.5 m) in the upper part. Scarce *Macoma simizuensis, Nuculana tumiensis, Glycymeris wishkahensis, Modiolus matchgarensis,* and *Lucinoma acutilineata* were found. The thickness is 30–32 m.

Sequence XI (Opoka-Aleurolitic)

24. Gray tuffaceous siltstones, slightly bedded and bearing coarse rubble interlayering with harder siliceous opoka interbeds (0.1–0.2 m), and clay interlayers, concretionary lenses, and siliceous horizons 0.2–0.3 m thick. Numerous tuff beds (0.1–0.2 m thick and

separated by intervals of 1.5 m) occur in the lower and middle parts of the member. Banded opoka beds (0.1 m) with alternating dark- and light-colored varieties predominate in the upper half. The sediments display microbedding and commonly contain small flakes of muscovite, rare pebbles, fine plant detritus, and scarce mollusks *Conchocele smekhovi, Macoma simizuensis, Chlamys* ex gr. *cosibensis,* and *Mytilus ochotensis.* The thickness is 38 m.

25. Similar bedded rocks, light-, dark-, and red-colored, hard, sometimes porcelaneous, bearing concretions (0.4 × 0.7 m) and loose ashy tuff interbeds (0.1 m thick). The deposits display coarse splinter parting, Liesegang rings, local banded structure, and jarosite-rich horizons up to 0.1 m thick. Gray siltstones, soft, slightly bedded, tuffaceous, enclosing three massive porcelaneous opoka beds bright yellow in color (0.3–0.5 m) predominate in the upper part of the member. The thickness is 25 m.

The total thickness of the Tumi Formation is about 480–490 m. The described deposits are overlain by a barren sequence conventionally assigned to the Pil' Formation. They are represented by slightly bedded, gray, coffee-colored clayey rocks bearing harder siltstone interbeds. The sediments are often breccated and composed of siltstone fragments (3–10 cm) concentrated in certain beds that are 0.2–0.3 m thick and cemented by a gray clayey matrix probably replacing the original tuffaceous material. On the other side of the Monchigar Lake, they appear to be in tectonic contact with tuffaceous diatomites of the Mayamraf Formation.

The Lower Reaches of the Vodopadnaya River (Basal Part of the Pil' Section)

Analogues of the Machigar and Tumi formations of the Machigar (northern) section are exposed along the Vodopadnaya River. Judging from separate outcrops, the former is represented by sandy conglomerates bearing carbonate concretions, boulders and pebble, coal lenses, and mollusks *Yoldia matschigarica, Liocyma furtiva, Chlamys matchgarensis,* and others. This unit is most likely up to several hundred meters thick. It is overlain by volcanic rocks represented by dark gray to greenish agglomerate lava breccia of trachybasalts containing fragments of amygdaloidal lavas and light-colored silicified sedimentary rocks. This sequence is evidently as thick as several tens of meters. Volcanites contain lenses (5 × 30 cm) of sedimentary bedded rocks. The lower contact of the sequence is unclear. It is overlain by sediments assigned to the Tumi Formation. Their basal part is represented by gray conglomerate bed (0.3–0.6 to 1 m), medium-pebbled, chiefly composed of basaltic pebbles and sporadical angular fragments. Pebbles are concentrated in the lower part of the sequence. The conglomerate bed rests conformably but with a hiatus. Upward the section, this bed is replaced by greenish gray sandstone (2.5 m), fine-grained, nonbedded, bearing numerous well-rounded pebbles. The overlying sediments are represented by ferruginate siltstone (5 m), displaying thin leafy parting and bearing pebble and thin (up to 3 cm) tuff interbeds. These rocks are in turn overlain by silty pebble-bearing sandstone (up to 10 m) grading upward into alternating sandstones and siltstones (16 m) with concretions (0.2–0.4 m), separate boulders, conglomerate interbeds (0.3–0.5 m), and shells of *Modiolus matchgarensis* and *Balannus* sp.

The overlying sediments are represented by opoka-like siliceous rocks (10 m), gray, brown at the surface, very hard, bedded, and bearing jarosite. They alternate with lighter yellow rocks (0.1 m), siltstone, clay (2–3 cm), and cream-colored concretionary interlayers (0.5 m). These rocks underlie a sequence of sandstones and siltstones (13 m) intercalated with bedded siliceous siltstones (0.2–0.5 m) and claystones with jarosite (up to 50–60 m thick).

The total thickness of the Tumi Formation is up to 100–115 m. Therefore, in the Pil' section, the Machigar and Tumi formations are represented by somewhat different facies and differ in thickness from the same units of the Machigar section. It is important that volcanic rocks missing in the northern section are confined here to the boundary between the studied formations.

MOLLUSKS AND SUBDIVISION OF THE SECTION

The analysis of the stratigraphic distribution of mollusks in the Machigar and Tumi formations revealed a succession of nine assemblages different in taxonomic composition (Fig. 2). This served as the basis for the subdivision of the studied formations into mollusk-bearing beds (I–IX), corresponding to stages of malacofauna evolution in the region and variations in paleogeographical, ecological, and facies environments.

In identifying these beds, we followed the procedure developed at the Geological Institute, Russian Academy of Sciences (Gladenkov, 1988). Lower boundaries of the beds were defined by the first occurence of new species. Certain beds were designated according to species names. In this case, preference was given to index species of one or another bed. When molluscan assemblages turned to be lacking index forms, we designated corresponding beds on the basis of newly appearing and dominant species. To characterize faunal assemblages we calculated the number of appeared, disappeared, endemic, cosmopolitan, and dominant species. We also used data on stratigraphic ranges of the studied molusks and on their ecological types.

When correlating our beds with stratigraphic units in adjacent regions, we focused attention on index species of a wide geographic range, on the successive appearance and disappearance of certain forms, and on the general composition of a given assemblage. In the case when analogues of certain beds are missing in adjacent regions or change their order, they have been

grouped into members that can be traced over a significant territory.

Characteristics of faunal beds of the Machigar section are presented below, together with indicated lithologic members and sequences, within which the beds were distinguished.

MACHIGAR FORMATION

Bed I with *Mytilus littoralis–Corbicula sitakaraensis–Cerithidea quadrimonilicosta* assemblage were distinguished in member 2 of the lower Machigar Subformation. The assemblage is relatively poor, containing only four shallow-water forms, three of which are marker species. Dominant are *Mytilus littoralis, Corbicula sitakaraensis,* and *Cerithidea quadrimonilocosta.*

Bed II with *Papyridea harrimani–Macoma sejugata* assemblage (members 3 and 4). The assemblage is rather poor, including 6 newly occurring species. Three forms, namely, *Portlandella watasei, Papyridea harrimani,* and *Macoma sejugata* pass into the overlying sediments. *Musculus* cf. *snatolensis* and *Acanthocardia* cf. *snatolensis* occur in this bed only.

Bed III with *Nemocardium iwakiense–Thracia schmidti* assemblage (lower part of member 5). The assemblage contains 15 forms, including three index species, i.e., *Nemocardium iwakiense, Thracia schmidti,* and *Clinocardium taracaicum.* The following species first occur in this bed: *Portlandella nitida, Liocyma furtiva, Modiolus matchgarensis, Yoldia caudata, Chlamys matchgarensis,* and *Periploma besshoensis.* Almost all of them are of a wide geographic range.

Bed IV with *Trachycardium kinsimarae–Yoldia matschigarica* assemblage (from the top of member 5 to the middle of member 9). The assemblage consists of 37 forms, including 7 index species, namely, *T. kinsimarae, Y. matschigarica, Arca sakamizuensis, Papyridea matschigarica, Colus asagaiensis, Nucula matschigarica,* and *Ainicardia napanica.* Twenty-five new species appear in the unit, among them *Ancistrolepis modestoideus, Pododesmus schmidti, Turritella importuna, Neverita asagaiensis, Yoldia laudabilis,* and *Ciliatocardium asagaiense. Portlandella watasei* and *Chlamys matchgarensis* make their last appearance in this bed. Representatives of genera *Ciliatocardium, Papyridea, Liocyma, Cyclocardia, Yoldia,* and *Mya* dominate in the assemblage. The bed yielded 5 endemics and over 30 species of a wide geographic distribution.

Bed V with *Acila oyamadensis* assemblage (the top of member 9–member 10). The assemblage contains 16 species, 13 of which passed from the underlying rocks. Relatively deep-water forms, including dominant *Periploma besshoensis, Yoldia caudata,* and *Macoma simizuensis,* make up the major part of the assemblage. The bed yielded one endemic form. The marker species is *Acila oyamadensis.*

Bed VI with *Glycymeris nakosoensis* assemblage (members 11–13). The assemblage of 24 species includes the index form *G. nakosoensis.* The last occurrences of *Macoma sejugata, Ciliatocardium asagaiense, C. yamasaki, Mya ceneiformis,* and *Ancistrolepis modestoideus* were recorded in the unit. Relatively deep-water species, including most frequent *Periploma besshoensis, Macoma simizuensis,* and *Conchocele smekhovi,* are still dominant in the assemblage. The bed contains 4 endemics and 20 forms of a wide geographic range.

TUMI FORMATION

Bed VII with *Delectopecten watanabei–Nuculana tumiensis* assemblage was distinguished in members 14–18 of the Tumi Formation. The assemblage contains 25 species, mostly inherited from the underlying deposits. Eleven species, including abundant deep-water forms *Delectopecten watanabei, Nuculana tumiensis, N. praerodbai,* and *Nucula tumiensis,* occur in the bed for the first time. The last occurrences of *Cyclocardia matschigarica, Turritella importuna, Neverita asagaiensis,* and *Macoma asagaiensis* were recorded at this level. The bed shows accumulations of mollusks *Periploma besshoensis, Conchocele smekhovi, Delectopecten watanabei,* and *Nuculana tumiensis.*

Bed VIII with *Cardiomya majanatschensis* assemblage (member 19 to the middle of member 21). The assemblage contains 15 species, 13 of them passed from the underlying sediments. The first occurrence of *Cardiomya majanatschensis* is recorded. The major part of the assemblage, consists of relatively deep-water species, whose representatives are *Nucula tumiensis, Nuculana tumiensis, Macoma simizuensis, Periploma besshoensis,* and *Conchocele smekhovi.* The bed yielded 2 endemics and 13 forms of a wide geographic range.

Bed IX with *Chlamys donmilleri–Mytilus ochotensis* assemblage were distinguished in the upper part of the Tumi Formation (middle part of member 21 to member 25). The assemblage contains 14 forms, including index species *Chlamys donmilleri, C. rikinensis, C.* ex gr. *cosibensis,* and *Mytilus ochotensis.* Species widely represented in the lower horizons of the Machigar and Tumi formations, such as *Periploma besshoensis, Modiolus matchgarensis, Nucula tumiensis, Conchocele smekhovi,* and others, almost completely disappeared from the bed. *Macoma simizuensis* is the most abundant form.

CORRELATION AND AGE

For comparison of the distinguished assemblages and beds, we used data on Paleogene mollusks from Sakhalin (Arkhipova *et al.,* 1992), Kamchatka (*Atlas ...,* 1984; Gladenkov, 1972; Gladenkov *et al.,* 1987), Koryak Upland (Volobueva and Krasnyi, 1979; *Atlas ...,* 1981), and Japan (Takeda, 1953; Kamada, 1962; Honda, 1989; Oyama *et al.,* 1960), which made possible the following correlation of beds I–IX. According

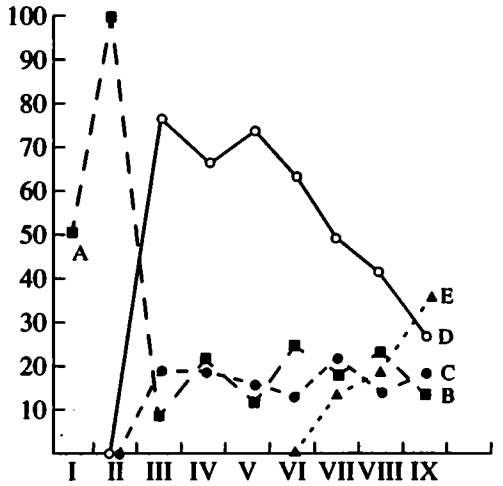


Fig. 3. Age groups of molluscan assemblages (%) in beds I–IX: (A) Eocene species; (B) Eocene–Oligocene species; (C) Oligocene species; (D) Oligocene–Miocene species; (E) Miocene species.

to the presence of *Corbicula sitakaraensis,* Bed I with *Mytilus littoralis–Corbicula sitakaraensis–Cerithidea quadrimonilicosta* assemblage can be conventionally correlated with the upper Eocene–Oligocene deposits of northern Japan, namely, with the Iwaki, Shitakara, and Shakubetsu formations. The occurrence of *Mytilus littoralis* suggests its correlation with the Eocene sediments of Sakhalin and Kamchatka, i.e., with the Krasnopol', Takaradai, and Snatol' formations. Bed II with *Papyridea harrimani–Macoma sejugata* assemblage is correlated with the upper Eocene–Oligocene sediments of northern Japan (Ishikari and Poronai formations, Urahoro and Ombetsu groups), Sakhalin (Takaradai, Arakai, and Gastello formations), and western Kamchatka (Kovachin and Amanin formations). Bed III with the *Nemocardium iwakiense–Thracia schmidti* assemblage and Bed IV with *Trachycardium kinsimarae–Yoldia matschigarica* assemblage bearing index species *Nemocardium iwakiense, Mya grewingki, Trachycardium kinsimarae, Papyridea matschigarica, Colus asagaiense,* and *Yoldia matschigarica,* together with newly occurring specimens of *Yoldia laudabilis, Y. caudata, Portlandella nitida,* and *Ciliatocardium asagaiense,* are correlated with the lower Oligocene sediments of northern Japan (Momijiyama, Asagai, Omagari, and Charo formations) and with Oligocene deposits of Sakhalin (Arakai and Gastello formations), Kamchatka (Amanin, Gakkh, Rategin, and Il'khatun formations), and Koryak Upland (Ionai and Malen formations). Beds III and IV evidently can be correlated with sediments of the Machigar Formation of the Pil' section, where the following mollusks were found: *Yoldia matschigarica, Liocyma furtiva, Thracia schmidti, Mya grewingki, Chlamys matchgarensis, Ancistrolepis modestoideus, Spisula* sp., *Solen* sp., *Clinocardium* sp., and *Macrocallista* sp. Bed V with *Acila oyamadensis* and Bed VI with *Glycymeris nakosoensis* contain index species that suggest their correlation with the Asagai Formation of the Joban coal field in northern Japan. The successive character of molluscan assemblages relative to mollusks of underlying beds III and IV and species *Portlandella nitida, Yoldia caudata, Y. laudabilis, Modiolus matchgarensis,* and *Pododesmus schmidti,* show that the beds can be correlated with the above-mentioned Oligocene formations from northern Japan, Sakhalin, and Kamchatka.

Beds VII and VIII in the lower and middle parts of the Tumi Formation contain index species *Delectopecten watanabei* and *Cardiomya majanatschensis,* in addition to mollusks inherited from the underlying deposits, such as *Modiolus matchgarensis, Turritella importuna, Yoldia laudabilis, Portlandella nitida, Periploma besshoensis, Conchocele smekhovi,* and *Macoma semizuensis.* Using these data, we correlated the beds with the Kholmsk Formation and basal beds of the Nevel'sk formations of Sakhalin, (the Makarov region), and also with the Utkholok–Vivintek Horizon of western Kamchatka, and accordingly dated them Oligocene in age.

Bed IX with *Chlamys donmilleri–Mytilus ochotensis* assemblage distinguished in the upper part of the Tumi Formation includes index species *Chlamys donmilleri, Chl. rikinensis, Chl.* ex gr. *cosibensis,* and *Mytilus ochotensis.* These species are known from the Lower Miocene deposits of Sakhalin (Chekhov Formation and upper Bora Subformation) and Kamchatka (Kuluven Formation). On this basis, we assign the bed to the lower Miocene.

Machigar and Tumi assemblages contain the Eocene, Eocene–Oligocene, Oligocene, Oligocene–Miocene, and Miocene species. Nine faunal beds distinguished in the section reveal a time succession of certain groups of mollusks (Fig. 3). For example, in the lower part of the section (Bed I of the Machigar Formation), the forms known from the Eocene and Eocene–Oligocene sediments of adjacent regions constitute almost 100% of the assemblage. The Bed II yielded the Eocene–Oligocene species. Assemblages from the beds III and VI are predominantly consist of Oligocene forms (63–75%), although they coexist with Eocene–Oligocene (25–8%) and Oligocene–Miocene (12–17%) species. The first Miocene mollusks constituting 13–16% of the assemblage appear in beds VII–VIII. However, the major part of the assemblage is represented at this level by Oligocene (43–47%) and Eocene–Oligocene (25–20%) species, whose mutual abundance is over 50%, whereas the Oligocene–Miocene mollusks are not abundant (16–20%) in beds VII–VIII. In Bed IX, age and compositions of molluscan fauna are considerably changed. Miocene (36%) and Oligocene–Miocene species (18%) are more numerous here than Eocene–Oligocene (18%) and Oligocene (27%) forms. Quantitative characteristics of molluscan assemblages, together with results of corre-

lation, validate our inference that places beds I–II in the upper Eocene and Eocene–Oligocene, beds III–VIII in the Oligocene, and Bed IX in the lower Miocene.

A comparison of these results with data on foraminifers, spores, and pollen suggests a similar correlation of the considered sequences. Previously, the following foraminifer units were distinguished in the Machigar section: *Trochammina matschigarica–Miliammina* sp. Beds spanning the coal-bearing part of the section, *Haplophragmoides laminatus–Pseudoelphidiella subcarinata* Zone of the Machigar Formation, and *Haplophragmoides carinatus–Porosorotalia tumiensis–Cyclammina pacifica–Islandiella curvicamerata* Zone of the Tumi Formation (Zhidkova *et al.*, 1981). In addition, the *Perfectononion praeincertus–Cribroelphidium sumitomoi* and *Reophax tappuensis* beds have been distinguished in the lower zone, and the *Rhabdammina aspera–Haplophragmoides subimpressus* Beds in the upper one. The studied foraminiferal assemblages enabled the correlation of coal-bearing deposits of the Machigar section with the Eocene sediments of California, southern Sakhalin, and Kamchatka; the Machigar Formation was correlated with the Eocene–Oligocene sediments of Sakhalin (Takaradai and Arakai formations) and western Kamchatka (Amanin, Gakkh, Utkholok, and Vivintek formations), and the Tumi Formation with the Oligocene–lower Miocene sequences of Sakhalin (Kholmsk and Nevel'sk formations). Beds and zones of the sequence in question correspond to the Machigar–Kholmsk stage of development of Cenozoic foraminifers in Sakhalin (Voloshinova *et al.*, 1970), whose assemblages are rather homogeneous in composition. Most species of this stage do not pass into overlying deposits. The stage is marked by a combination of Paleogene forms and newly occurring Miocene species.

The Machigar and Tumi formations, excluding the uppermost part of the sequence, bear the Central Tamlev assemblage of spores and pollen. On the basis of palynomorphs, the studied sediments were correlated with the Gastello, Kholmsk, and Nevel'sk (basal bed) formations of southeastern Sakhalin, with the Amanin, Gakkh, Utkholok, and Vivintek formations of western Kamchatka, and with the Il'khatunvayam Formation of the Karaginskii Island in northeastern Kamchatka (Arkhipova *et al.*, 1992). In the upper part of the Tumi Formation and at the base of the Pil' Formation from the Pil' section, there was detected the Berezov spore–pollen assemblage, suggesting that deposits are correlative with the lower Miocene sequences, such as the Kuluven Formation of western Kamchatka, the *Laternula* Sandstone of northeastern Kamchatka, and the Uina Formation and upper Nevel'sk Subformation of eastern Sakhalin.

PALEOECOLOGICAL, PALEOCLIMATIC, AND PALEOBIOGEOGRAPHICAL ENVIRONMENTS

Molluscan fauna from the described beds of the Machigar section shows a gradual transition from lagoonal brackish-water assemblages to shallow marine, and further to relatively deep-water mollusk population areas outside the shelf zone.

The lower unit of the section distinguished as Bed I yielded the *Mytilus–Corbicula–Cerithidea* assemblage characteristic of littoral or upper sublittoral brackish-water settings with a hard substrate. Next, the *Musculus–Acanthocardia–Portlandella–Macoma* assemblage with thin-walled species is confined to the clayey layers of Bed II. It evidently populated a quiet lagoon with normal salinity.

Mya–Nemocardium and *Papyridea–Liocyma–Ciliatocardium* assemblages indicative of upper sublittoral environments, occur in beds III–IV, where disjoined and broken valves are mixed with intact shells, suggesting a reasonably active hydrodynamic regime of bottom waters.

Beds V and VI are most rich in relatively deep-water species of the *Periploma–Macoma–Conchocele–Portlandella–Yoldia* assemblage, evidently residing in the lower sublittoral area. Scarce, separate, and sometimes broken valves of relatively shallow-water forms *Glycymeris, Peronidea,* and *Musculus* most likely indicate their posthumous transportation from a shallow to a deeper sublittoral zone.

The molluscan assemblage from beds VII–VIII is of particular interest. It is dominated by representatives of *Delectopecten, Nuculana,* and *Nucula* genera, associated with less abundant *Conchocele, Periploma, Macoma,* and *Portlandella* forms. Large accumulations of dominant species suggest that the depth of paleobasin was greater than 200 m. The observed combination of transported broken shells of shallow-water shelf genera *Ciliatocardium, Modiolus, Pododesmus, Glycymeris* and deep-water species shows these assemblages to have been buried on the continental slope.

In the uppermost Bed IX of the Tumi Formation, deep-water forms became scarce and are replaced by the *Macoma–Nuculana* assemblage. The occurrence of numerous casts and molds of *Chlamys* species at this level may indicate the initial and gradual shoaling of the basin.

Molluscan assemblages from the Machigar and Tumi formations contain both relatively cold-water forms and taxa whose modern distribution areas are located to the south of northern Sakhalin. Data on current geographic ranges of mollusks occurring in Oligocene deposits of the Schmidt Peninsula show that 13 of 44 recorded genera inhabit the lower latitudes, namely, northern Japan, Primor'e, and southern Sakhalin, down to the latitude of the Terpenie Cape. This implies that the Oligocene marine climate was considerably warmer than at present. However, the bed-by-

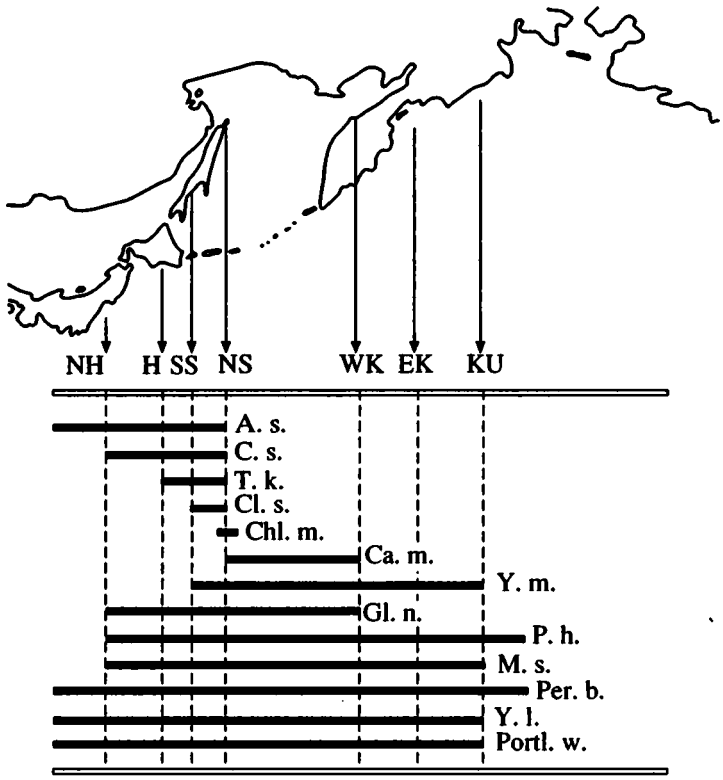


Fig. 4. Distribution areas of index and dominant molluscan species from the Machigar and Tumi formations in the northwestern Pacific: (A. s.) *Arca sakamizuensis;* (C. s.) *Corbicula sitakaraensis;* (T. k.) *Trachycardium kinsimarae;* (Cl. s.) *Ciliatocardium schmidti;* (Chl. m.) *Chlamys matchgarensis;* (Ca. m.) *Cardiomya majanatschensis;* (Y. m.) *Yoldia matschigarica;* (Gl. n.) *Glycymeris nakosoensis;* (P. h.) *Papyridea harrimani;* (M. s.) *Macoma sejugata;* (Per. b.) *Periploma besshoensis;* (Y. l.) *Yoldia laudabilis;* (Portl. w.) *Portlandella watasei.* (NH) northern Honshu; (H) Hokkaido; (SS) southern Sakhalin; (NS) northern Sakhalin; (WK) western Kamchatka; (EK) eastern Kamchatka; (KU) Koryak Upland.

bed analysis of fauna in the section shows the discrete character of stratigraphic distribution of thermophilic species. For example, beds I, III, IV, VI, VIII, and IX can be assigned to relatively warm epochs marked by the appearance of thermophilic genera, whereas the intervals of beds II, V, and VII correspond to colder periods. Using the benthic fauna as a basis, it is difficult to judge whether this is the direct influence of paleoclimatic changes or the result of depth variations in the paleobasin. The data on spores and pollen (Zhidkova *et al.*, 1981; Arkhipova *et al.*, 1992) show that the Machigar period and greater interval of the Tumi time were characterized by a rather stable climate. The single Central Tamlev spore–pollen assemblage was distinguished in the respective interval of the section. Substantial changes in vegetation are recorded only in the upper part of the Tumi Formation, where the Berezov spore–pollen assemblage is of a more thermophilic type. In our opinion, this part corresponds to Bed IX with the *Chlamys donmilleri–Mytilus ochotensis* assemblage.

The studied geographical distribution of mollusks recorded in the Machigar and Tumi formations revealed several groups of species typical of different distribution areas (Fig. 4): forms known from the Oligocene of northern Japan and Sakhalin, Sakhalin endemics, species reported from the Oligocene of Sakhalin and western Kamchatka, and Oligocene taxa characteristic of northern Japan, Sakhalin, and western Kamchatka. A considerable part of the studied assemblages is represented by species of a wider distribution area extending from northern Japan and Sakhalin to northeastern Kamchatka and Chukotka. They can be classified as the North Pacific species. According to the distribution areas, the quantitative biogeographic characteristics of molluscan assemblages are as follows. The Machigar assemblage as a whole includes 24% Sakhalin–Northern Japanese species, 14% Sakhalin endemics, 2% West Kamchatka–Sakhalin–North Japanese forms, and 60% of North Pacific taxa. Among mollusks from the Tumi Formation, we recorded 13% of Sakhalin–North Japanese forms, 9% Sakhalin endemics, 9% Sakhalin–West Kamchatka species, 13% Sakhalin–West Kamchatka–North Japanese taxa, and 56% North Pacific species. The results show that species of a wide geographic range are dominant in both the Machigar and Tumi assemblages. This indicates a steady connection between marginal marine basins of the northwestern Pacific, from Japan to Chukotka, during the Oligocene time.

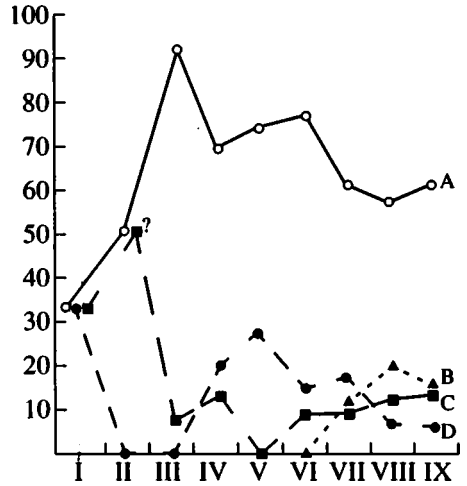


Fig. 5. Abundance of biogeographical groups of mollusks (%) in beds I–IX: (A) North Pacific species; (B) Sakhalin–West Kamchatka and Sakhalin–West Kamchatka–North Japanese species; (C) Sakhalin species; (D) Sakhalin–North Japanese species.

On the other hand, mollusks of different distribution areas replacing each other in the meridional direction most likely suggest the past climatic differentiation in the northwestern Pacific and coexistence of species somewhat different from the viewpoint of thermopathy.

The above discussion shows that a considerable number of species of a wide geographic range (over 50%) first appeared in the Machigar Formation only in Bed III, corresponding to the onset of normal marine environments in the region, and then continued to exist up to the top of the Tumi Formation. The underlying beds I and II contain only a few North Pacific forms of a wide geofraphic range. Moreover, a considerable number of endemic Sakhalin species (33% and more, Fig. 5) suggests a certain isolation and endemic character of faunas in shallow lagoons during the terminal Eocene and initial Oligocene.

CONCLUSION

A detailed study of mollusks from the Paleogene–Neogene sediments of the Machigar and Tumi formations of the Schmidt Peninsula in northern Sakhalin resulted in the distinction of nine molluscan beds and provided the basis for their correlation with coeval units in adjacent regions of northern Japan, Sakhalin, Kamchatka, and Chukotka. On the basis of correlation, the section of the Schmidt Peninsula was subdivided into upper Eocene (the lower coal-bearing part of the Machigar Formation), Oligocene (Machigar and Tumi formations, excluding the uppermost part of the latter), and lower Miocene (upper horizons of the Tumi Formation) intervals.

The paleoecologic analysis of molluscan assemblages in the section made it possible to trace the gradual transition from lagoonal deposits at the base of the Machigar Formation to normal marine shelf sediments (Machigar Formation) and further to slope deposits of the Tumi Formation.

The study of paleoclimatic changes indicates that the Oligocene marine climate of northern Sakhalin was warmer than at present. According to spore–pollen assemblages, the warmest interval of the studied section was defined in the upper part of the Tumi Formation (early Miocene). Paleobiogeographical records suggest a steady connection between marginal seas of the northwestern Pacific in the Oligocene and early Miocene time.

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The Holocene Palynostratigraphy and Geochronology of Lacustrine–Palustrine Deposits in the Kola Peninsula

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Abstract—The studied objects represent the lacustrine bottom deposits in the southeast of the Kola Peninsula and the peat bed in the Cape Rybachii, exemplifying the northernmost occurrence of this substance in the European part of Russia. A detailed radiocarbon dating and palynological analysis provided grounds to establish the Holocene chrono- and climatostratigraphy of the region and to correlate the studied sections with coeval deposits in Scandinavia. Sedimentation in the lakes commenced in the region 11 000–10 000 years ago. The pine species spread over the Kola Peninsula 9000–7000 years ago. The peat accumulation was initiated here 8000–7500 years ago. The birch tundra forests advanced up to the Barents Sea coast 6000–5000/4500 years ago. The northern forest boundary retreated southward after the climate deteriorated 4500 years ago. When the pine forest areas decreased, peat bogs became more widespread in the southeastern Kola Peninsula, and the tundra belt appeared at the coast of the Barents Sea. Spruce trees arrived at their northern distribution limit in the peninsula about 3500 years ago. The current structure of the vegetation cover was formed during the last 2500–2000 years.

Key words: *Kola Peninsula, Holocene, radiocarbon dating, palynology, chronostratigraphy.*

Studies on the Holocene deposits of the Kola Peninsula were initiated in the 1930s (Poretskii *et al.*, 1934); however, even at present, only few spore–pollen diagrams from this region are substantiated by the detailed dating results (Lebedeva, 1977; Elina and Lebedeva, 1982; Vashchalova and Klimanov, 1987; Elina *et al.*, 1995). The convincing dating of three lacustrine and palustrine section (Fig. 1) allows us to detect the precise time succession of the landscape evolution in the peninsula during the almost entire Holocene and to correlate these sections with coeval deposits in Finland, Sweden, and Norway. We calculated our noncalibrated radiocarbon dates using the ^{14}C half-life constant equal to 5568 years.

PEAT BED IN THE CAPE RYBACHII (69°38′ N, 32°22′ E)

A section of the upland moor in the Cape Rybachii was studied by S.V. Goryachkin and A.E. Cherkinskii in 1990. The moor is located northward of the Gulf of Eina in a small relief depression at an altitude of 60 m above sea level. It represents the northernmost peat locality in the European part of Russia dated at present. The radiocarbon dating of molluscan shells from the surface of marine terraces of the cape showed that the peat bed remained above the sea level throughout the Holocene (Koshechkin *et al.*, 1976).

The Cape Rybachii is connected with the Kola Peninsula through the Srednii Isthmus. Basement rocks of the Cape Rybachii are represented by the Upper Proterozoic conglomerates, sandstones, and shales (*Atlas ...*, 1971). Quaternary deposits consist here of glacial beds accumulated during the last glaciation. The lichen–fruticulose to fruticulose–lichen tundra is the principal element of the cape landscape. In addition, there are populations of *Betula* sect. *Nanae* in small river valleys at the southern coast. The birch elfin woodlands are also known in the opposite continental coast of the Motovskii Gulf, separating the Cape Rybachii from the Kola Peninsula; the distance between them and the upland moor in question is about 25 km.

The peat bed is 110 cm thick. The radiocarbon dating for the bed (Table 1) was performed at the Radiometric Laboratory of IGRAS. Dates obtained, except the rejuvenated date IGRAS-1590, omitted in our interpretation, represent the regular succession. According to the dating results, the peat formation commenced about 7500 years ago. K.V. Kremenetskii was responsible for the analysis of spores and pollen from the peat samples, and O.N. Uspenskaya investigated their botanical composition.

The structure of the peat bed is as follows:

0–10 cm: wet sphagnous upland peat composed of *Sphagnum lindbergii* Schimp. and *S. nemoreum* Scop.

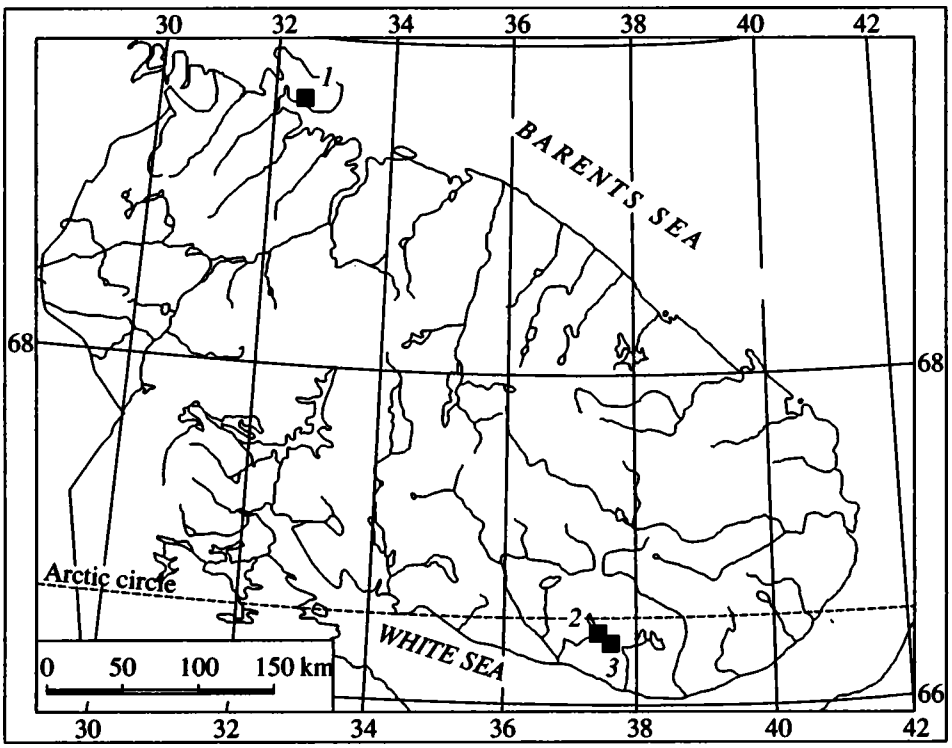

Fig. 1. Geographic position of studied sections: (1) Cape Rybachii, 60°38′ N, 32°22′ E; (2) Babozero Lake, 66°22′–66°34′ N, 37°18′–37°34′ E; (3) Krugloe Lake, 66°22′ N, 37°35′ E.

in association with accessory hypnous *Drepanocladus fluitans* Warnst. and sedge *Carex magellanica* Lam.

10–20 cm: arboreal–sphagnous transitional peat consisting of *Sphagnum lindbergii* Schimp., birch wood and bark, hypnous *Drepanocladus* sp., and sedge species *Carex* sp. and *C. magellanica* Lam.

20–50 cm: sphagnous upland peat with *Sphagnum nemoreum* Scop. representing the main component and associated with *Sphagnum teres* Aoengstr., *S. warnstorfii* Russ., and sedge *Carex* sp.

Table 1. Radiocarbon dates for peat samples from the Cape Rybachii, 60°38′ N, 32°22′ E

Depth, cm	Age, years ago	Sample no.	Material
0–10	628 ± 43	IGRAS-1433	Peat
10–20	1629 ± 37	IGRAS-1396	"
20–30	3514 ± 38	IGRAS-1587	"
30–40	3528 ± 36	IGRAS-1434	"
40–50	4425 ± 40	IGRAS-1588	"
50–60	5233 ± 67	IGRAS-1589	"
60–70	6143 ± 42	IGRAS-1435	"
90–100	4030 ± 255	IGRAS-1590	"
100–110	7438 ± 64	IGRAS-1436	"

50–100 cm: arboreal–herbaceous lowland peat composed of foliaceous wood mixed with birch and willow bark; at a depth of 50–60 cm, the peat mass includes alder bark, indeterminant herbs, sedge remains, true and sphagnous mosses, and horsetail *Equisetum* sp.

100–110 cm: herbaceous lowland peat with indeterminant herbs, willow bark and wood, reed *Phragmites australis* (Cav.) Trin. ex Steud., sedge *Carex* sp., and horsetail *Equisetum* sp.

The spore–pollen diagram obtained (Fig. 2) revealed three palynozones.

The lower Palynozone 1 corresponds to the time interval of 7500–5000/4500 years, i.e., to the Atlantic period of the Holocene. Pollen of arboreal and fruticose taxa, mainly that of birch and, to a lesser extent, of pine and willow, preponderate in the general composition of palynomorphs. We also detected pollen of alder at the level approximately corresponding to the time mark of 6100 years ago. Pollen of herbaceous group generally consists of specimens of Gramineae and Cyperaceae. This group also includes Rosaceae, *Menyanthes trifoliata* L., and forbs. Spores are produced by fern and horsetail species.

Spectra of arboreal and fruticose pollen are variable. The pine pollen content is at its minimum in the lower sample and grows upward, as is evident from the pertinent variation curve in the diagram. The peak abun-

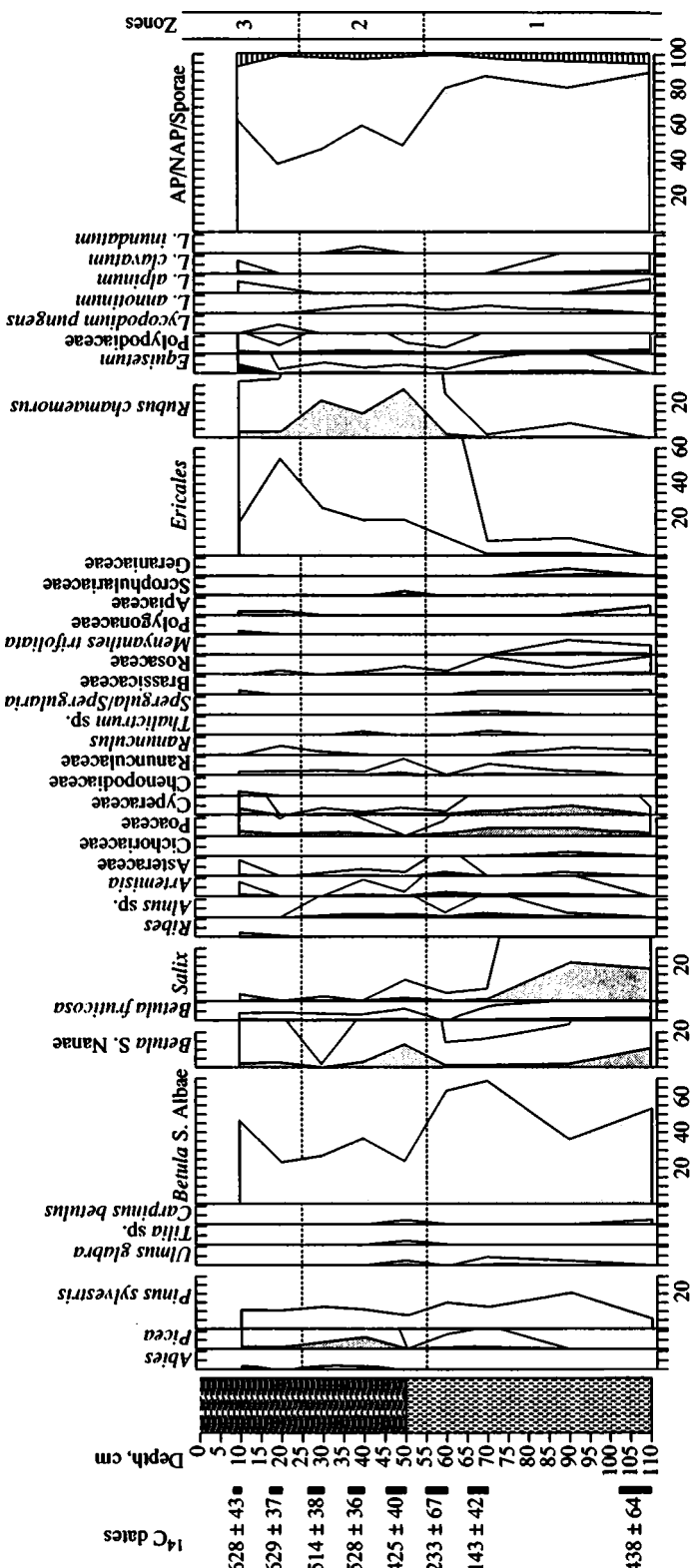

Fig. 2. Spore–pollen diagram for peat samples from the Cape Rybachii, 60°38′ N, 32°22′ E (analyst K. V. Kremenetskii).

Table 2. Radiocarbon dates for lacustrine sediments from the southeastern Kola Peninsula

Babozero Lake, 66°22′–66°34′ N, 37°18′–37°34′ E			
Depth, cm	Age, years ago	Sample no.	Material
90–100	3090 ± 180	Beta-55324	Humic acids
190–200	9430 ± 280	Beta-55325	"
290–300	126 ± 400	Beta-55326	"
Krugloe Lake, 66°22′ N, 37°35′ E			
Depth, cm	Age, years ago	Sample no.	Material
80–90	3670 ± 90	Beta-55329	Humic acids
190–200	6440 ± 80	Beta-55330	"
290–300	7310 ± 80	Beta-55331	"
385–395	8040 ± 90	Beta-55332	"
490–500	8970 ± 110	Beta-55333	"
533–543	10000 ± 220	Beta-55334	"

dance of willow pollen is recorded in the lower part of Palynozone 1. In the upper half of this palynozone, the willow pollen content abruptly decreases, and this is compensated by a simultaneous rise in the variation curve of birch pollen.

The middle Palynozone 2 corresponds to the time interval of 5000/4500–3500 years, in other words, to the Subboreal period of the Holocene. The general composition of palynomorphs is enriched here in herbaceous pollen and is consequently depleted in arboreal and fruticose taxa (Fig. 2). The interval is marked by sharp rises in the variation curves of Ericaceae and *Rubus chamaemorus* pollen. Simultaneously, there is a drop in the birch pollen distribution curve, and the pine pollen abundance becomes somewhat lower. A small peak in abundance of the pine specimens is recorded at the level of about 4000 years ago. The pollen of alder is a permanent component, but all ecdemic broad-leaved taxa are missing in this interval.

The radiocarbon dating indicates that the peat accumulation ceased between 3500 and 1700 years ago.

The upper Palynozone 3 corresponds to the last 1700 years, i.e., to the Subatlantic period of the Holocene. The distribution curve of cloudberry pollen dips down in this palynozone, whereas alder pollen disappears here. In contrast, Ericaceae and horsetail taxa are in peak abundance. The upper sample from this palynozone showed the maximum of the birch pollen abundance, and, as a result, the total content of arboreal and fruticose taxa also increased at this level (Fig. 2).

LAKES IN THE SOUTHEASTERN KOLA PENINSULA

Bottom deposits of two lakes in the southeastern Kola Peninsula have been studied in 1992 by the expedition of the Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Russian Academy of Sciences (IGOPMG RAS), headed by N.G. Patyk-Kara. Other participants of the fieldwork were P. Kolinvo (Smithsonian Tropical Institute of Miami, USA), M. Miller (University of Cincinnati, USA), O.N. Pakhomova and E.V. Tikhomirova (IGOMPG RAS), V.V. Kol'ka (Institute of Geology, Kola Scientific Center), and V. Bakhmutov (Institute of Geophysics, Ukrainian Academy of Sciences). Spores and pollen from the samples were analyzed by L.V. Golubeva. The radiocarbon dating (Table 2) was performed in the Beta Laboratory, Miami, Florida, USA, and supported by grant of P. Kolinvo.

Basement rocks in the southeastern Kola Peninsula are represented by Archean and Proterozoic gneisses, schists, and amphibolites (*Atlas ...*, 1971). The Quaternary sedimentary cover is thin and consists of glacial and glacial–lacustrine deposits of the Valdayan glaciation. A strip of kame moraine with knob-and-kettle topography is located southward of the Babozero and Krugloe lakes. The area to the east of the Babozero Lake is very boggy. Altitudes in this region of the peninsula are in the range of 200–300 m.

Lakes are widespread in the Kola Peninsula (Kagan *et al.*, 1992). Sedimentation in the lakes was initiated soon after the glacier retreat. The area under discussion is located in the northern taiga subzone, and lakes here are usually surrounded by bogs. The current spruce–birch and pine–birch forests mainly belong to the *Lichen–Bryales* and *Lichen–Picea–Betula* types.

The Babozero Lake is one of the largest in the Kola Peninsula, being 23 km long and 1–3 km wide. It is located between 66°22′–66°34′ N and 37°18′–37°34′ E. The lake area is over 5000 hectares (Fig. 3). The lake is 2–3 m deep on average. Bottom sediments were drilled in the southern stretch of the lake at about 66°22′30″ N and 37°32′ E. At the sampling site, its depth is 1.5 m. We recovered here the core column of bottom deposits 3.2 m long and consisting of the following beds:

Bed 1, subbottom interval of 0–70 cm: dark brown diatomaceous gyttja;

Bed 2, subbottom interval of 70–230 cm: laminated gyttja (alternating black and brown laminae) with abundant vivianite inclusions in the lower portion of the interval;

Bed 3, subbottom interval of 230–270 cm: homogeneous brown diatomaceous gyttja; and

Bed 4, subbottom interval of 270–320 cm: coarse-grained to gravely sand with an admixture of light brown clay and abundant mica flakes.

The spore–pollen diagram for the Babozero section includes five palynozones (Fig. 4). The lower part of this lacustrine section (Bed 4) is barren of spores and pollen.

Pollen content in Bed 3 is insignificant, and the lower Palynozone 1 comprises this unit and the lower-

most sample from Bed 2. This palynozone corresponds to the time interval from 11 000 to 10 000 years ago. As compared to other intervals of the section, Palynozone 1 is the least rich in arboreal and fruticose pollen taxa, but it shows the highest abundance of herbaceous pollen and spores. Fruticose and fruticulose birch taxa, chiefly *Betula nana*, are dominant in the first taxonomic group. Gramineae, Cyperaceae, and Ericaceae preponderate among herbaceous pollen, whereas ferns are most abundant in the spore group. This palynozone is also marked by a rather high abundance of *Lycopodium alpinum* spores.

Palynozone 2 is correlated with the time interval of 10 000–8000 years (the terminal pre-Boreal and Boreal periods of the Holocene). In this palynozone, the variation curve of total content of arboreal and fruticose taxa rises, whereas the distribution curves of spores and herbaceous pollen go down. This trend in variation of spore–pollen spectra remains valid higher in the section. The arboreal–fruticose pollen group demonstrates here an abrupt increase in abundance of pine pollen and decreasing content of *Betula nana* pollen. Pollen of alder is not abundant. In addition, the variation curve of spruce pollen goes up at the level of the upper sample of Palynozone 2.

Palynozone 3 is recognized in the upper part of Bed 2. Lacustrine deposits of this palynozone were accumulated between 8000 and 3300/3000 years ago, i.e., in the Atlantic and initial Subboreal periods of the Holocene. This palynozone is distinct because of the gradually rising curve of spruce pollen distribution, and also due to abundance maxima of pine (up to 51%) and alder pollen, which are more prevalent than elsewhere in the section. At the same time, the content of fruticose birch pollen is at its minimum in this interval. In contrast to other palynozones, this one is also richer in pollen of aquatic plants.

Palynozone 4 corresponds to the lower part of Bed 1, deposited 3000–1000 years ago (at the end of Subboreal time and in the first half of Subatlantic period of the Holocene). In its interval, the percentage of spruce pollen in the spore–pollen spectra is maximal. Abundance of pine pollen is somewhat lower at this level, whereas the fruticose birch taxa are more prosperous than before. Pollen of alder is missing here, but specimens of Ericaceae are still present. Simultaneously, the total content of spores slightly increases mainly because of a higher abundance of sphagnous forms, whose concentration, as compared to other spore–pollen spectra, is highest.

The upper Palynozone 5 corresponds to the second half of the Subatlantic period (the last 1000 years). This palynozone marks a decrease in the total content of arboreal and fruticose pollen taxa and an increase in abundance of herbaceous pollen. The latter becomes richer in the Gramineae and Cyperaceae, whereas the percentage of sphagnum spores goes down. At the same time, pollen grains of dwarf, fruticose, and arbuscular

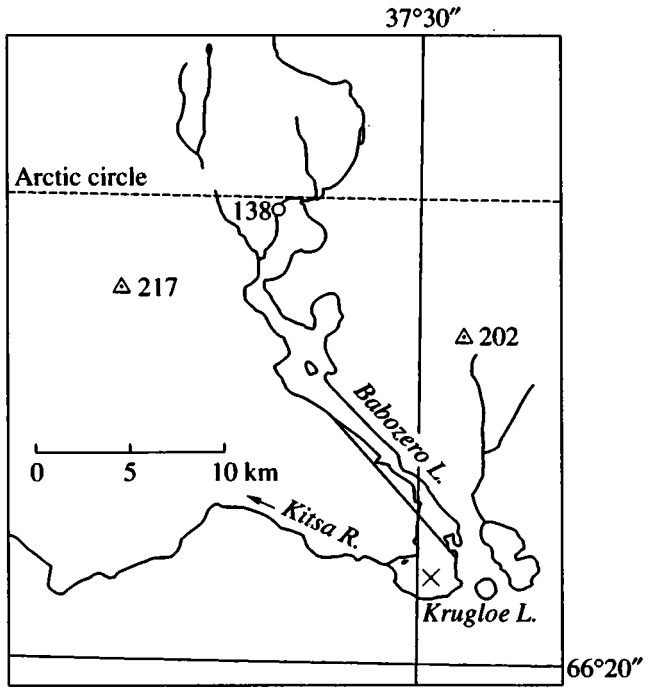

Fig. 3. Geographic scheme of Babozero and Krugloe lakes with indicated drilling site (cross) and elevation marks (triangles).

birch taxa are much more plentiful here than in other spectra, whereas the proportion of coniferous palynomorphs is lower.

Lake Krugloe (66°22′ N, 37°35′ E, altitude 140 m) is situated in a small depression near the southeastern coast of Lake Babozero. It is round-shaped, with a diameter of about 1 km and a total area of about 80 hectares (Fig. 3). The lake is 1–2 m deep on average, and the depth at the sampling site is 1 m. Bottom sediments have been penetrated by drilling down to 5.5 m, and their section is of the following structure:

Bed 1, subbottom interval of 0–485 cm greenish brown diatomaceous gyttja;

Bed 2, subbottom interval of 485–545 cm dark gray mud with abundant vivianite;

Bed 3, the apparent subbottom interval of 545–550 cm coarse-grained sand with an admixture of light brown clay.

The radiocarbon dating of basal lacustrine loam suggests that sedimentation in the lake commenced about 10 000 years ago at the level of the Pleistocene–Holocene boundary.

The Lake Krugloe section is more comprehensively analyzed and calibrated by dating. Moreover, this lake is much lesser in size, which is favorable for using its spore–pollen diagram for the Holocene chronostratigraphy of the southeastern Kola Peninsula (Fig. 5).

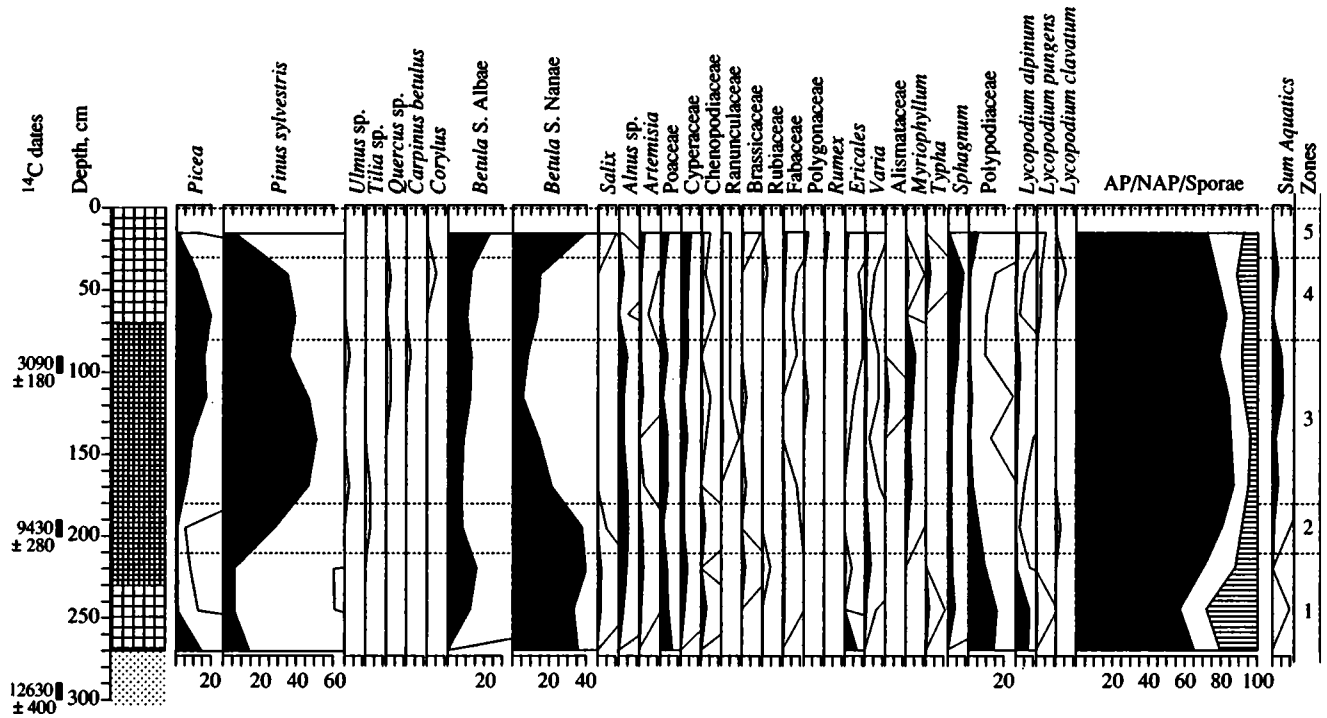

Fig. 4. Spore–pollen diagram for bottom sediments from the Babozero Lake, 66°22′–66°34′ N, 37°18′–37°34′ E (analyst L.V. Golubeva).

Pollen of arboreal and fruticose taxa makes up 80 to 90% of total amount of counted spores and pollen grains, and it thus is dominant in all spore–pollen spectra throughout the section. The percentage of herbaceous pollen can be as high as 20%, and that of spores is up to 10%. Approximately half of all pollen grains represents fruticose birch taxa.

The lower Palynozone 1 of the section characterizes Bed 2 and corresponds to the pre-Boreal and initial Boreal periods of the Holocene, comprising a time interval of 10 200–9000 years ago (Fig. 5). It differs from other palynozones by the lowest (less than 80%) content of arboreal pollen, and consequently, by the highest (up to 20%) concentration of herbaceous taxa.

The pollen of *Betula* species, especially of *Betula nana* (up to 39–42%), is predominant component of the arboreal pollen group of Palynozone 1. The pollen of coniferous taxa is not abundant. Herbaceous plants are represented by pollen of Gramineae, Cyperaceae, and forbs. Ferns preponderate in the sporiferous group, where they associate with Boreal and tundra club-mosses (*Lycopodium pungens, L. alpinum,* and *L. clavatum*).

Palynozone 2 is detected in the lower part of Bed 1. Judging from results of radiocarbon dating, it corresponds to the Boreal period of the Holocene ranging from 9000 to 8000 years ago. The general composition of pollen spectra remains constant at this level without significant changes. However, variation curves display here a somewhat rising abundance of pine and arbuscular birch pollen, as well as a decreasing content of *Betula nana* pollen. In addition, the spore–pollen spectra in question are distinct because of appearance of the alder species *Alnus incana* (L.) Moench. in their composition.

Changes in the taxonomic composition of pollen and spores are more significant in Palynozone 3. According to radiocarbon dates, this palynozone spans the time interval of 8000–7000 years ago. Pollen of arboreal and fruticose taxa is more abundant here (up to 80–90% in sum), and, consequently, the content of herbaceous pollen is lower. These proportions between main components of spore–pollen spectra are also characteristic of higher intervals of the section.

Pine (up to 32% of all pollen grains) is dominant among other arboreal taxa of Palynozone 3. Other characteristic features of this zone are the considerably lower abundance of arboreal birch pollen, a higher content of fruticose birch pollen, and a somewhat greater proportion of alder pollen than before.

Palynozone 4 corresponds to the greater interval of the Atlantic period, ranging from 7000 to 5000/4500 years ago. This palynozone is marked by a rise in the variation curve of spruce pollen. In the upper part of the interval of 6000–5000 years ago, we detected the peak abundance of pine pollen and minimum content of *Betula nana* pollen. The proportion of sphagnous spores also grows here. The sedimentation rate in the lake during the last 5000/4500 years abruptly decreased as compared to the time interval of 9000–5000 years ago.

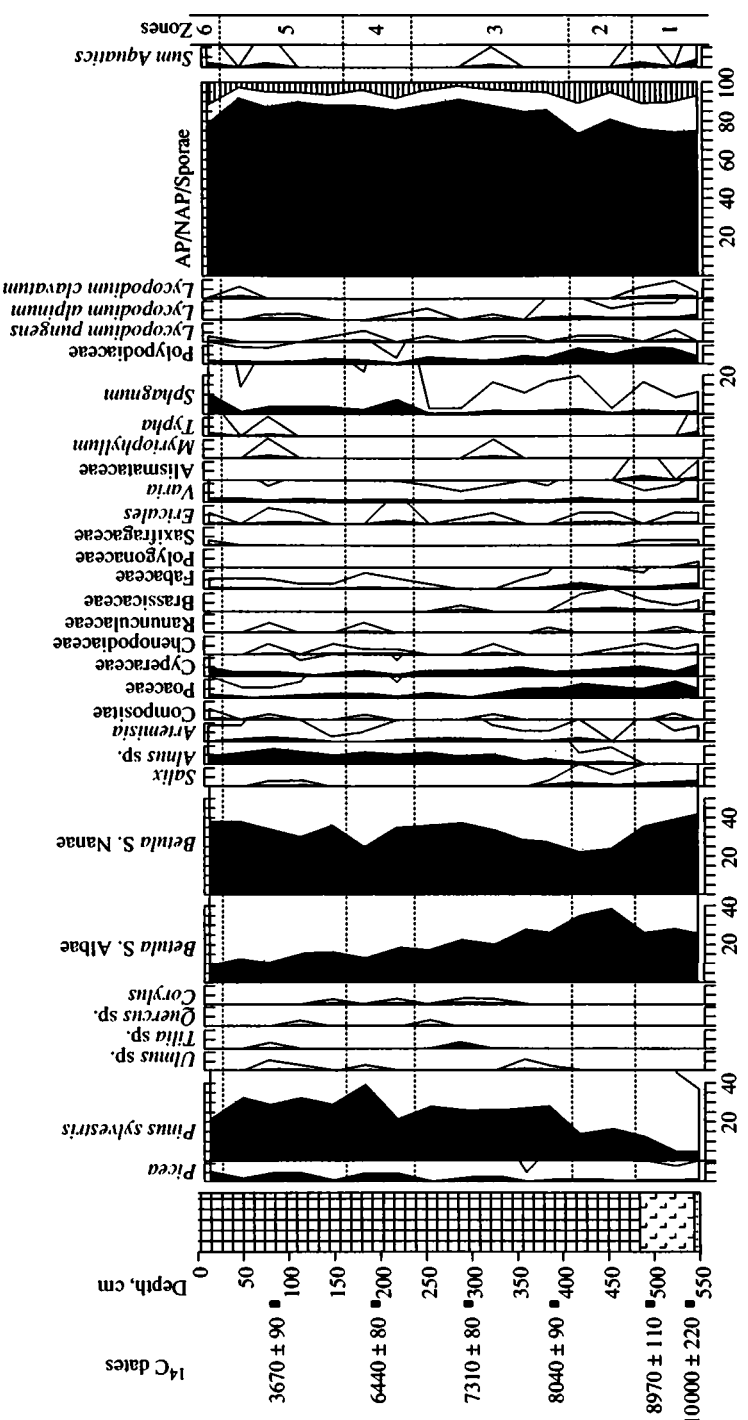

Fig. 5. Spore–pollen diagram for bottom sediments from the Knugloe Lake, 66°22′ N, 37°35′ E (analyst L.V. Golubeva).

Palynozone 5, correlative with the Subboreal period of the Holocene (5000/4500–2500 years ago), differs from the underlying zone only by minor changes in the composition of spore–pollen spectra. It is slightly depleted in pine pollen and richer in fruticose birch pollen than the upper part of Palynozone 4.

The uppermost sample of the section (Palynozone 6 comparable with the Subatlantic period that commenced 2500/2000 years ago) showed that the total content of arboreal pollen decreased to 78%, whereas the summary percentage of herbaceous pollen and spores grew mainly due to a higher abundance of sphagnous spore taxa (Fig. 5).

DISCUSSION

Our results of radiocarbon dating, spore–pollen, and botanical analysis, when compared with published palynological diagrams, have enabled us to elucidate the Holocene chronostratigraphy of the Kola Peninsula and to correlate events in this and adjacent areas in northern Finland.

Available radiocarbon dates suggest that close to 11 000 years ago environments in the southeastern Kola Peninsula were rather mild and favorable for the appearance of tundra landscape with fruticose and dwarf birch species around the Krugloe and Babozero lakes. The birch forest-tundra with accessory willows existed near the lakes approximately 10 000 years ago. The last landscape also included areas of dwarf birch-and-moss tundra.

Pine appeared in the southeast of the Kola Peninsula about 9000 years ago, when the habitat areas of arbuscular birch also expanded. Birch taxa were most widespread in forests during the time period of 9000–8000 years ago. The appearance of alder pollen at the level corresponding to 9000 years ago shows that the distribution area of this tree approached the Kola Peninsula at that time. Other components of tundra vegetation did not change.

The period between 8000 and 7000 years ago corresponded to the expansion stage of forest areas, and their birch population was substituted to a certain degree by pine trees. The alder species *Alnus incana* (L.) Moench. became a member of forest communities in the southwestern Kola Peninsula about 7500 years ago.

At the beginning of the Atlantic period of the Holocene (about 7500 years ago), i.e., at the commencement of peat accumulation in the lowland bog of the Cape Rybachii, the northwestern Kola Peninsula was occupied by tundra landscapes with accessory birch taxa. Willows were widespread in the thicket areas, and both types of trees coexisted in bogs. About 7000 years ago, pine migrated up to the current limit of its distribution in the northwestern Kola Peninsula.

Spruce species *Picea obovata* Ledeb. arrived in forest communities of the southeastern Kola Peninsula approximately 7000 years ago. It is likely that pine and mixed forests have been most widespread in the region from 6500 to 5000 years ago. This was a period of optimum climatic conditions. Pine and spruce forests were abundant in the region throughout the Atlantic period of the Holocene. The presence of ecdemic pollen grains of broad-leaved arboreal taxa suggests that their geographic ranges have approached the Kola Peninsula in this period. Similar data were obtained for Karelia as well (Elina, 1981).

Bog formation around Lake Babozero and general expansion of sphagnous bog areas were activated about 6500 years ago, as it follows from the local peak abundance of sphagnum spores in Palynozone 4. During the thermal optimum period 6000–5000 years ago, the upper forest boundary in the Khibiny Mountains could be 250–300 m higher than at present (Vashchalova and Klimanov, 1987). At the time of about 5500/5300 years ago, spruce species *Picea obovata* Ledeb. advanced into the Lovozero region located to the northeast of the Khibiny Mountains (Elina *et al.*, 1995).

The general composition of vegetation in the Cape Rybachii was relatively constant between 7000 and 5000/4500 years ago. Nevertheless, the botanical composition of the peat shows that the abundance of fruticose willow taxa abruptly decreased in the peat bog community of plants around 6000 years ago. Birch thickets in the cape and within the studied peat bog were most widespread from 6000 to 5000/4500 years ago. At this time, the birch forest-tundra zone was spread as far to the north as the Barents Sea coast.

Despite the low concentration of alder pollen in the studied samples, this tree was definitely present in the Cape Rybachii, because its botanical remains were detected in the peat. Judging from palynological data, we may assume that alder has appeared in the Cape Rybachii about 6100 years ago. At present, the nearest habitats of alder are located in the Pechenga River valley 40 km to the southwest of the bog locality in the Cape Rybachii and in the coastal zone of the Kola Gulf (*Flora* ..., 1956). The former migration of alder to the north was in response to favorable climatic conditions in the second half of the Atlantic period.

The Atlantic–Subboreal boundary corresponding to the time mark of 4500 years ago was most important in the second half of the Holocene history of the Kola Peninsula. Climatic changes of that time resulted in cooling and more intense rainfall in the Khibiny Mountains (Vashchalova and Klimanov, 1987). The temperature decrease down to the minimum was between 4500 and 4100 years ago. This event was responsible for the formation of permafrost localities under peat bogs around the Lovozero Lake and in some areas of the Khibiny Mountains (Vashchalova, 1986; Elina *et al.*, 1995). Afterward, the peat bog of the Cape Rybachii was covered with sphagnous mosses and entered the upland stage of evolution. Birch trees (*Betula tortuosa* Ledeb.) survived only in local areas of the Cape Rybachii, and tundra landscapes with cloudberry and

Ericaceae became widespread. The tundra zone also spread over the Barents Sea coast. In corresponding diagrams, the ecdemic pollen of broad-leaved arboreal taxa is missing, and this suggests that the northern boundary of broad-leaved forests in Karelia and Finland has retreated backward (Elina, 1981).

About 3500 years ago, spruce species *Picea obovata* Ledeb. arrived at the northern limit of its current distribution area in the northwestern Kola Peninsula. Alder vanished from the Cape Rybachii not later than 3500 years ago.

By 3000 years ago, the pine forest areas in the southeastern Kola Peninsula decreased, and spruce turned out to be more abundant in local forest communities. Sphagnous peat bogs became widespread.

The first half of the Subatlantic period (approximately 1700–1000 years ago) is marked by the abruptly lowered abundance of cloudberry in the herbaceous tundra communities of the Cape Rybachii. This time also stimulated *Sphagnum lindbergii* Schimp. to be preponderant among other bog mosses. Brushwood of Ericaceae became most widespread in the regional landscape.

CONCLUSION

Data on the studied sections revealed similar trends in the evolution of tundra and northern taiga zones of the Kola Peninsula during the Holocene. When the region became free of glaciers, it was covered first with tundra vegetation, and forest communities migrated in this territory later, during the climatic warming at the beginning of the Holocene.

Birch taxa were first to populate the new territory. Somewhat later, they were followed by pine trees, whose appearance was important for the forest formation in the region. The speckled alder was next. Spruce was the last of arboreal taxa that entered the Kola Peninsula, where it occupied the most favorable habitat areas and intensively replaced pine.

The late Atlantic climatic optimum is well manifested in different areas of the Kola Peninsula. In this period, the birch forest–tundra spread up to the coast of the Barents Sea. Data for adjacent regions of Finland (Mäkelä *et al.*, 1994; Seppla, 1996) also show that the frontal boundary of forest vegetation has been located here during this period farther north than at present. In addition, this warming phase is recorded in the mountain regions of Sweden and Norway (Moe and Odlund, 1992; Kullman, 1995). Thus, all the data of paleoecology and available models of secular climatic fluctuations substantiate the conclusion that global warming in the northern latitudes happened about 6000 years ago (Foley *et al.*, 1994).

The late Atlantic climatic optimum was followed by the climate's deterioration less than 4500 years ago. General environmental changes in the tundra and northern forest zones obviously were under influence of global climatic fluctuations of the Holocene time (Velichko *et al.*, 1994). The Subboreal and Subatlantic periods of the second half of the Holocene stimulated the bog formation in the region, which was especially intense during the last 1000 years. Forests survived in those areas of the southeastern Kola Peninsula that had better drainage systems, for instance, in the moraine ramparts and in areas covered with soil of a light mechanical composition (fluvioglacial sands). Natural environments and landscapes of the Kola Peninsula resembling the current situation initially appeared in the Subatlantic period.

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Principal Stages in the Geologic History of Georgia in the Paleogene

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Abstract—Paleogeographic and lithofacies schemes were compiled on the basis of a study of the Paleogene deposits, consideration of their thickness, and estimation of drilling data. The geologic history of Georgia in the Paleogene is presented. The width of the Paleogene flysch and epicontinental (the Ksani–Arkaly paraautochthon) basins in the southern slope of the Greater Caucasus, which they had prior to folding and thrust deformation, was evaluated. As a result, the extent of reduction of the basins was established. Formations of three types, namely flysch, epicontinental–marine, and volcanogenic–sedimentary sequences, accumulated in different geodynamic environments (marginal sea, island arc, and interarc rift) during different time intervals of the Paleogene. The Pyrenean tectogenesis (the Paleo- and Neopyrenean phases of folding) resulted in significant paleogeographic reorganization in Georgia, as in the entire Caucasus region. These phases marked the onset of the orogenic stage and created major morphostructures that subsequently controlled the modern tectonic patterns in the region. Time range and character of manifestation of the Neopyrenean folding phase are established, and general regularities in development of the Pyrenean tectogenesis in Georgia are considered.

Key words: *Paleogene, paleogeography, lithologic schemes, Georgia, epochs of tectogenesis.*

Georgia occupies a part of the Caucasus folded region. It shows complicated tectonic patterns and no less complicated geologic history.

The Paleogene, especially its second half (the late Eocene–Oligocene), represents one of the revolutionary stages in the geologic history of the Caucasus and the whole Alpine fold system.

An elucidation of peculiar features of geologic structure and evolution, along with a compilation of paleogeographic and lithofacies schemes for separate epochs of the Paleogene, made it possible to distinguish regular changes of depositional conditions in time and space. This is especially important for regions where deposits are buried under younger formations or overlapped by allochthonous sheets.

The Paleogene deposits occur to a variable extent in all structures, with the exception of the anticlinorium of the Greater Caucasus (Fig. 1). The deposits display more complicated tectonic patterns within the fold system of the southern slope of the Greater Caucasus, where their exposures are encountered in the Gagra–Dzhava and Mestia–Tianeti zones. In the nappe structure of the Mestia–Tianeti zone, it is nonetheless difficult to reveal the true zoning of lithofacies.

Sections along the Aleura and Ksani rivers where the Paleogene deposits are represented more completely have been used to restore original dimensions of Paleogene basins in this tectonic zone.

The performed lithofacies and structural analyses have resulted in palinspastic reconstructions of Paleogene basins of the eastern area of the southern slope of the Greater Caucasus. The width of basins prior to folding, as well as their subsequent reduction because of folding and nappe formation, have been evaluated. It has been found that the degree of reduction for flysch deposits of the Zhinval–Gombor subzone in the Mestia–Tianeti zone averages 45–50%, while that for subplatform deposits (the Ksani–Arkaly paraautochthon) is up to 25–30%. The overall amplitude of thrusting for sheets that overlap the Paleogene deposits on the southern slope is up to 9–12 km.

Borukaev (1970) has made palinspastic reconstructions for a western flysh basin within the southern slope of the Novorossiisk synclinorium. According to his data, the coefficient of reduction for the trough during the thrusting stage amounts to 2.8–2.9.

From viewpoint of Gamkrelidze (1976), the reduction for the Adzharia–Trialeti fold zone constitutes 25% of the regional territory.

A large body of works recently dedicated to the consideration of the geological history of the Caucasus has been based on ideas of mobilism. From recent standpoints, the territory of Georgia in the earliest Paleogene represented a part of the northern active margin of the Tethys similar to the western periphery of the Pacific Ocean, where a system of island arcs, as well as marginal and interarc basins, is developed. An island arc and marginal sea of the Greater Caucasus, together with North and South Transcaucasian island arcs, separated by the Adzharia–Trialeti inter-arc rift, were dis-

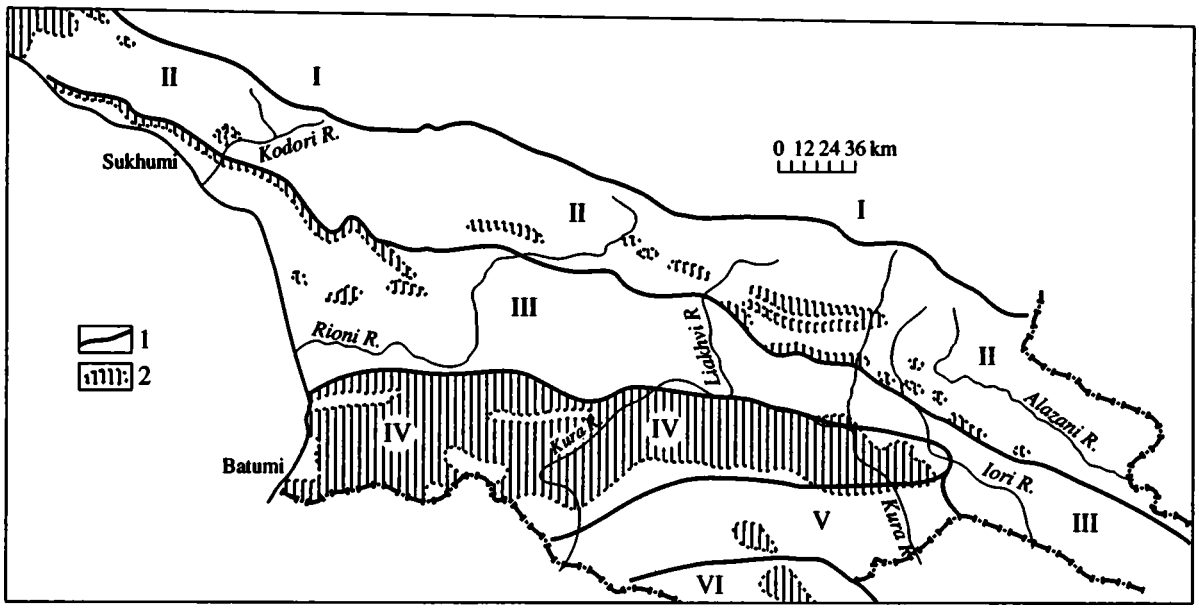

Fig. 1. Simplified scheme of the tectonic zones in Georgia (after Gamkrelidze and Gamkrelidze, 1978): (1) boundary of tectonic units; (2) natural exposures of the Paleogene deposits; (I) anticlinorium of the Greater Caucasus; (II) fold system of the southern slope of the Greater Caucasus; (III) Georgian Block; (IV) Adzharia–Trialeti zone; (V) Artvin–Bolnisi Block; (VI) Loksk-Karabakh zone of folding.

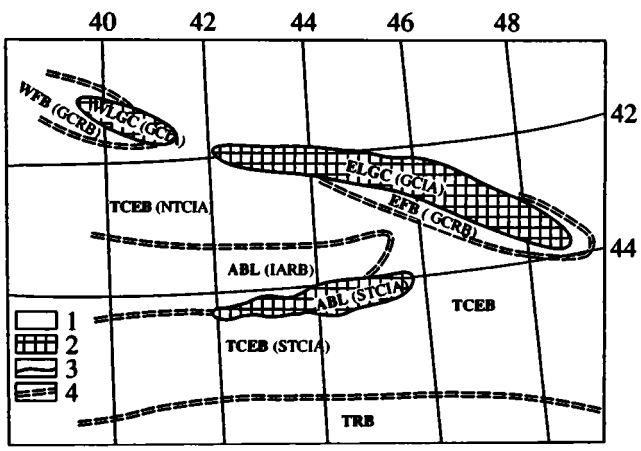

Fig. 2. Palinspastic scheme of the Caucasian sector in the northern margin of the Tethys for the earliest Paleogene (compiled with due regard to paleomagnetic data and palinspastic schemes by Adamia *et al.*, 1979; Lortkipanidze *et al.*, 1984; Gamkrelidze, 1986): (1) sedimentation area; (2) erosion area; (3) boundary of dry land; (4) boundary of sedimentation basins with different environmental conditions; (WLGC) Western Land of Greater Caucasus; (ELGC) Eastern Land of Greater Caucasus; (WFB) Western flysch basin; (EFB) Eastern flysch basin; (TCEB) Transcaucasian epicontinental basin; (ATB) Adzharia–Trialeti basin; (ABL) Artvini-Bolnisi land; (TRB) Tethyan relic basin; (GCIA) Greater Caucasus island arc; (GCRB) Greater Caucasus remnant basin; (IARB) inter-arc rift basin; (NTCIA) North Transcaucasian island arc; (STCIA) South Trancaucasian island arc.

tinguished in our region, as main structural units of the Cretaceous–Paleogene boundary time (Adamia *et al.*, 1991).

In my opinion, somewhat different principal morphostructures existed here by the beginning of the Paleogene (Fig. 2).

The respective basins included small islands that played a significant and sometimes decisive role in the formation of facies zones. There were also the Rachinsk–Vandam cordillera zone and the Tkibuli–Sachkhere, Trialeti, Gurian, and other land areas (Fig. 3).

The Rachinsk–Vandam cordillera zone, bordering the Eastern flysch basin of Greater Caucasus on the south, was of particular significance. The zone consisted of separate cordilleras and extended along the southern boundary of the flysh basin from the Utsery in the west inside the northern periphery of the Gagra–Dzhava zone, which is now completely overlapped by allochthonous thrust sheets of the Cretaceous–Paleogene flysh deposits. The Rachinsk–Vandam cordillera zone was a main provenance of terrigenous material for the Paleogene flysch basin. The composition of clastics in the upper Eocene olistostromes on the southern slope of the Greater Caucasus indicates that the cordillera zone was mainly composed of the Mesozoic and, partly, of Paleogene rocks of the Gagra–Dzhava zone. Beginning from the Aragvi River eastwards, it also included pre-Jurassic rocks of the crystalline basement (Maisadze, 1994).

Among morphostructures of sedimentation areas, there were submarine uplifts, whose role in formation

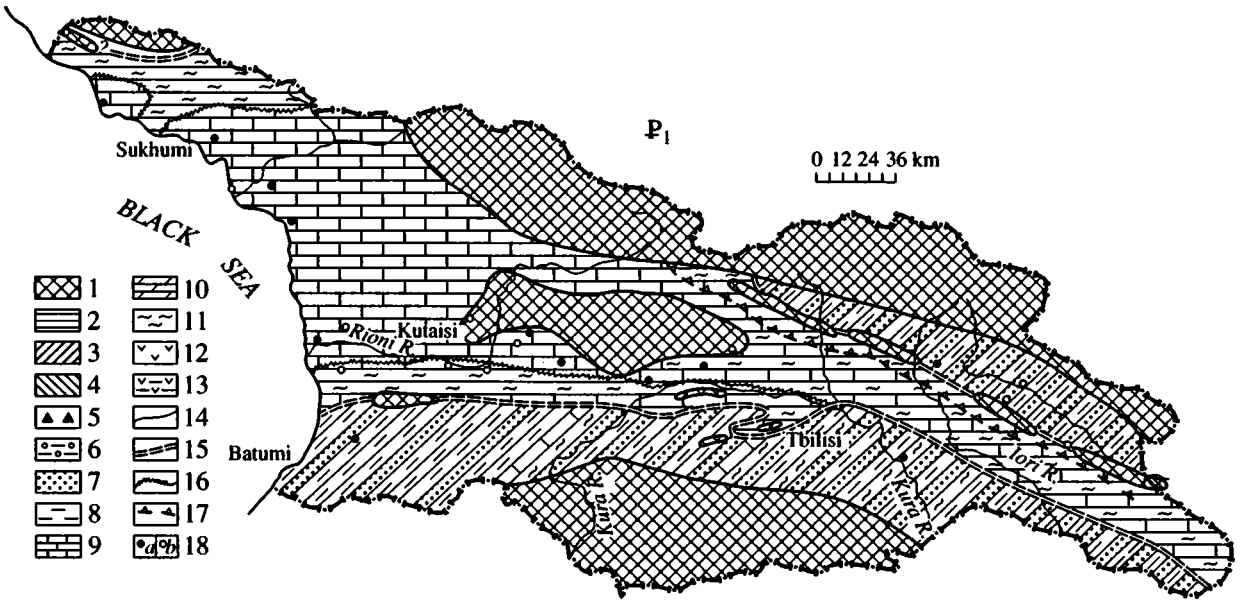

Fig. 3. Paleogeographic and lithofacies scheme of Georgia for the Paleocene (compiled by Maisadze): (1) dry land; (2) normal sedimentary formation; (3) flysch and flyschoid formation; (4) volcanogenic–sedimentary formation; (5) olistostrome; (6) conglomerate, breccia, gritstone; (7) sandstone, siltstone; (8) clay; (9) limestone; (10) dolomite; (11) marl; (12) mostly volcanic tuff and lava; (13) alternation of volcanogenic and normal sedimentary rocks; (14) boundary between sea and land; (15) boundary between formation zones; (16) boundary between lithofacies; (17) present position of flysch overthrust; (18) drill holes: (*a*) deposits of a given age present, (*b*) deposits of a given age are missing.

of lithofacies zoning is often underestimated. Submarine uplifts in the vicinity of Gagra and Novyi Afon may be an example. The uplifts separated the epicontinental basin into shallow troughs, which evolved subsequently into isolated depressions with different sedimentation regimes (the eastern margin of the Adler depression and the Gudauta and Ochamchire depressions).

Some submarine uplifts represented areas of the basin with "zero sedimentation." In my opinion, areas of present-day Gumista and Kodory rivers and regions near the Tsebelda Village exemplify such uplifts. Gavtadze (1986), who studied nanoplankton, points out that, in some sections, an upper part of the middle Eocene deposits occurs directly on the lower Eocene beds without any sign of unconformity. At the same time, in areas adjacent to the uplifts, sections of fossiliferous Paleogene deposits are complete, lacking hiatuses.

During the Paleogene (except for the Oligocene), the Western and Eastern flysch basins of Greater Caucasus represented accumulation areas of respective sedimentary formations. Between these two deep-water troughs adjacent to the Greater Caucasus Islands in the north, a shallow-water basin with the epicontinental sedimentation regime probably existed in the present Bzyb–Kodori interfluve area. The basin was similar to that located in the Georgian Block and Northern Caucasus (Fig. 3). Such an assumption is based on the lithofacies analysis of the Paleogene deposits of the two regions. This assumption is also confirmed by the comparative analysis of fauna from the Paleocene–Eocene deposits of the Georgian block and Northern Caucasus (Salukvadze, 1993). A similar paleogeographic situation probably existed north of the Great Caucasus Land, as well, within the Cis-Caspian–Kuban trough (Sarkisyan *et al.*, 1987).

The North Transcucasian island arc situated southerly (south of the flysch basins) was covered by an epicontinental sea that accumulated thick terrigenous–carbonate deposits. A similar sedimentation regime existed in the Artvin–Bolnisi block during the Early Paleogene and in the Loksk–Karabakh zone at the end of the early Eocene. Flysch (Paleocene–lower Eocene), as well as volcanogenic–sedimentary (middle Eocene and partially upper Eocene) and carbonate–terrigenous (Upper Eocene–Oligocene) deposits, accumulated in the Adzharia–Trialeti rift basin with uniform facies conditions, unlike the other areas under consideration.

Thus, three types of formations (flysch, volcanogenic–sedimentary and epicontinental–marine) accumulated in Georgia under dissimilar geological conditions (a marginal sea, an island arc, an intra-arc rift) at different times.

As is evident from recent data on lithofacies of the Paleogene deposits and also from the analysis of their thickness, published materials, and drilling data, the principal stages in the Paleogene geologic history of Georgia can be outlined as follows.

In the Paleocene (Fig. 3), sediments accumulated probably in the Western flysch basin of the Greater

Caucasus (the upper courses of Psou and Bzyb rivers), the eastern termination of the Novorossiisk basin, although this conclusion is based on insufficiently plausible data. West of the basin (in the Chvezhipsa zone), the Plastunskaya Formation accumulated at that time (Keller and Menner, 1945).

The continuous sedimentation was still in progress in the Eastern flysch basin of the Greater Caucasus (east of Bolshaya Liakhvi River). Clastic–limestone and siltstone–sandstone flysch 22–25 m thick (the Kvetera Formation) was deposited here in the Danian. In the late Paleocene, the carbonate sedimentation slowed down, and only siltstone–sandstone flysch 100–260 m thick accumulated (the Shakhvetili Formation).

A shallow inland sea has existed since the beginning of the Paleocene in the Transcaucasian epicontinental basin that occupied the most extensive area. Bedded, in places coarse-bedded limestone accumulated in this sea together with marl beds confined to some parts of the basin (the eastern part of the Adler depression and in the Otkhara–Apsta interfluve area).

Small islands that had existed by the end of the Cretaceous were completely flooded in the course of the Paleocene transgression (near the Mt. Tskakhaya and Shkhemi, Sachikobavo, and some other localities).

Further east, within the Kvaisi ore region, where the Paleocene and lower Eocene fossiliferous deposits were found some time ago (Maisadze, 1986), pelitomorphic clayey limestones and sandy marls accumulated at that time.

Similar paleogeographic and facies conditions existed in the Adzharia–Trialeti rift basin during the Paleocene and early Eocene, and its geologic history is considered below, together with the epicontinental sea basins.

As a result of the Danian regression, marls and clays of the variegated formation accumulated, instead of the Upper Cretatious limestones.

The regression was followed by a transgression, which was responsible for an increase of sedimentation areas. As a consequence of differentiated tectonic movements, a central zone of submergence was formed, which was later to became an area where thick flysh (Paleocene–lower Eocene) and volcanogenic–sedimentary deposits (middle Eocene) accumulated.

This zone was bounded by cordilleras and submarine uplifts on the north and south. West of the Abastumani, the zone was divided by transverse submarine uplifts into two parts. Thick (1500–3000 m) flysch deposits of the Borzhomi Formation accumulated east of the Abastumani in a deep-water trough. Sediments transitional between flysh and subplatform deposits and similar to those penetrated by boreholes in the Chakhati anticline (the Kintrish River basin) presumably accumulated west of the Abastumani (with the exception of the Guriiskaya depression). Judging from several cores samples, these deposits (1000 m thick) are of the carbonate–terrigenous flyschoid type (Nadareishvili, 1981).

The Paleocene and lower Eocene marls and clays of the Variegated Formation accumulated at the periphery of the rift basin and on submarine uplifts.

The localized volcanic activity that resulted in outflow of dacitic lava took place on the southern slope of the Teleti Range. Individual outbursts of volcanism occurred at the end of the early Eocene, as well, when the "Volcanogenic Flysch" (the upper part of the Borzhomi Formation), i.e., terrigenous–volcanogenic deposits, accumulated in the central part of the rift basin.

The problem of provenance for the terrigenous deposits of the flysch basin is still controversial. Judging from the remarkable length of the flysch trough (up to 260 km) and remarkable the thickness of flysch deposits (1500–3000 m), it is evident that a vast land area under erosion is required to be a provenance for the extra- and intraclastic resedimentation. According to recent opinion, the terrigenous material for the flysch basin was derived from the Georgian and Artvino–Bolnisi blocks, as well as from the eroded cordilleras located within the basin. The cordilleras undeniably supplied the flysch basin with terrigenous material. However, if we take into account that their dimensions have decreased due to the Paleocene transgression (especially of the Trialeti cordillera) and that the Variegated Formation mainly accumulated in the proximity of the cordilleras, then the erosion areas inside the basins can hardly be considered as the basic source of the terrigenous material. As for the Georgian Block, which bounds the rift basin in the north, it accumulated epicontinental deposits such as marls and clays (similar to rocks of the Variegated Formation in the northern part of the Adzharia–Trialeti zone) practically throughout the Paleocene–early Eocene period. Therefore, in my opinion, the assumption that the flysch trough received material from the erosion area in the Georgian block is invalid.

It is evident from the above reasoning that the basic portion of turbidity flows has arrived to the flysh basin from the Artvino–Bolnisi block located in the south. Intraformational breccias occurring in flysch deposits near the Algeti River are composed of rewashed crystalline rocks of the Artvino–Bolnisi block, and this partially lends support to the aforesaid theory (Gamkrelidze, 1976).

The early Eocene (Fig. 4) flyschoid deposits, which are composed of sandstone, sandy limestone and marl with rare interlayers of fine-clastic conglomerates, were accumulated in the western flysch basin (the upper courses of Psou and Bzyb rivers).

The siltstone–sandstone flysch sequence (a lower part of the Kvakeriskhevi Formation) 60–75 m thick

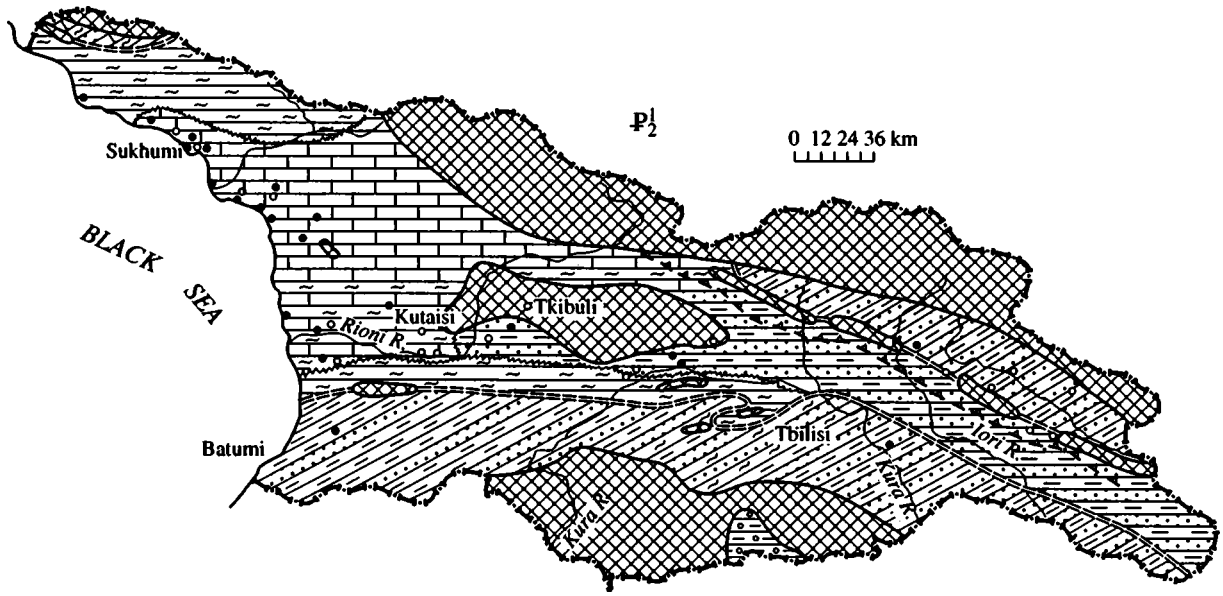

Fig. 4. Paleogeographic and lithofacies scheme of Georgia for the early Eocene (compiled by Maisadze; symbols as in Fig. 3).

accumulated during this time in the eastern flysch basin (east of the Bolshaya Liakhvi River).

Limestone and marl accumulated under carbonate sedimentation conditions in the epicontinental basin located further south.

Deposition of variegated marls with limestone interlayers (25–55 m) was characteristic of the eastern part of the Adler depression throughout the early Eocene. The situation was similar in the Bzyb and Apsta interfluve area. Shallow-water limestone with marl interlayers accumulated in many areas of the Georgian Block. Inflow of the terrigenous material took place east of the Tkibuli–Sachkheri block, where carbonate sandstone were deposited.

Conglomerate, sandstone and sandy limestone accumulated in the southeastern part of the basin (the Loksk–Karabakh zone) due to transgression in this area late in the early Eocene.

The middle Eocene (Fig. 5) accumulation of flyschoid deposits was still in progress in the western flysch basin. Sedimentation environments were here as follows: psammitic flyschoid deposits without admixture of carbonates were formed in the western part of the basin (the Psou River upper stream), while sandy clastic limestone and carbonaceous arkosic to graywacke–quartzite sandstone accumulated in the eastern part (the Bzyb River upper stream).

Psammitic silty flysch (190–200 m thick), which makes up the upper part of the Kvakeriskhevi Formation, accumulated during the early Eocene in the eastern flysch basin. In some sections (the Aleura and Pshavskaya Aragvi rivers), crystal and ashy trachyandesitic tuffs, as well as tuffaceous mudstone interbeds are present in the middle Eocene layers of this formation. These volcanites up to 2 m thick are lens-like.

The transgression in the epicontinental basin that started in the Paleocene peaked in the middle Eocene. Sea covered the most part of the Rachinsk–Vandam cordillera zone, where accumulation of condensed sandstone, limestone, and marl sequence began.

The transgression transformed some islands in the Satandzhio, Rechkhi, Zugdidi, and Tskhakaya areas into sedimentation setting. Marls with syngenetic pyrite were deposited under reduction conditions in the eastern margin of the Adler depression and in many areas of Abkhazia and Mingrelia. The area of the Novyi Afon submarine uplift was an exception; the sea shallowed here and nummmulitic limestone accumulated. Clayey limestone was deposited at that time within the Rachinsk–Lechkhumi syncline and Kvaisini region.

Against the backgrounds of the terrigenous–carbonate sedimentation, there was inflow of volcanic ash from the Adzharia–Trialeti zone (the rift basin) into some areas of the epicontinental basin. This inflow was responsible for deposition of thin beds of crystal and ashy tuffs basic to intermediate in composition (Nokalakeev anticline and villages of Achandara, Gachedili and Chkvishi). Some volcanic ash reached also the flysch basin on the southern slope of the Greater Caucasus (Aleura and Pshavskaya Aragvi rivers).

The middle Eocene in the Adzharia–Trialeti rift basin was marked by important events indicating a new stage in its geologic history. As a result of crustal extension, rifting resumed in the central zone of subsidence. This triggered an intensive volcanism first in the western part and then in the eastern part of the basin. The

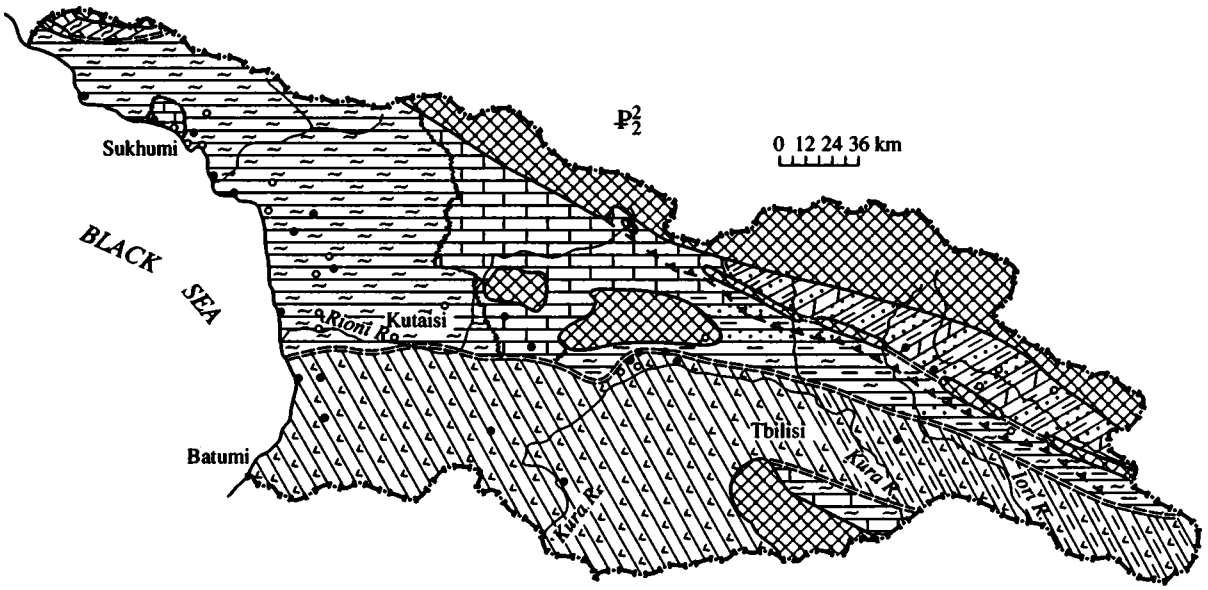

Fig. 5. Paleogeographic and lithofacies scheme of Georgia for the middle Eocene (compiled by Maisadze; symbols as in Fig. 3).

transgression was progressing during the basin subsidence.

Rifting, volcanism, and transgression predetermined the peculiar features of the geologic history of the Adzharia–Trialeti zone in the middle Eocene.

Intense downward movements transformed the central zone of subsidence, especially its western part, into a deep-water trough, where volcanogenic–sedimentary deposits up to 7 km thick accumulated during the middle Eocene. On the basis of petrochemical composition of volcanic rocks, Lortkipanidze (1980) distinguished the western and eastern segments of the trough, where vertical and transverse zoning was established.

The western segment includes three rock complexes corresponding to three basic stages of the middle Eocene volcanism. The rocks are represented by lava and volcaniclastics of tholeiitic hornblende and olivine basalts, basaltic andesites, andesites, potassic and subalkalic basalts.

In the eastern segment, synchronous complexes consist of low-K basalts (the axial zone), shoshonites (the northern zone), and subalkalic basalt (the southern zone), thus displaying the transverse zoning. According to Nadareishvili (1981), three transverse volcanic segments corresponded to three structural segments distinguished in the Adzharia–Trialeti zone on the basis of manifestations of the Paleogene and Cretaceous volcanism (Gamkrelidze, 1976). The transverse segments differ in the nature of volcanism and distribution of lithofacies, and also in the dynamics of sedimentation.

At the end of the middle Eocene, the Trialetian (Old Paleopyrenean) phase of folding in the Adzharia–Trialeti zone resulted in the partial relief inversion and sea regression. As a consequence, the central part of the basin, which was an area of intense sedimentation until this moment, turned into a zone of uplifts that separated the Paleogene basin into the northern and southern parts (Fig. 6).

Owing to differential movements along deep-seated faults (Chokhatauri, Surami–Gokishuri), the Gurian depression originated in the middle Eocene.

Peculiar deposits known as "intricately bedded conglomerates" appeared near Tbilisi (Abikh, 1870). Judging by their genesis, these deposits represent typical mixtites which I recommend to name the Tbilisian Olistostrome. In my opinion, tectonic processes at the end of the middle Eocene, i.e., in the period of maximum activity of the Trialetian orogeny, played a leading part in the formation of this olistostrome.

The middle Eocene transgression spread over the main part of the Artvini–Bolnisi Block, which, with the exception of a small area (Khram and Loki crystalline salients), was covered by sea. Thick (2–2.5 km) volcanogenic–sedimentary sequence that had accumulated here consisted of agglomerate breccias and basaltic lavas associated with layered andesite–dacitic to liparite–dacitic tuffs and rhyodacitic ignimbrites.

It is likely that an epicontinental sea, where limestone and marl accumulated, existed east of the Khram and Loki blocks, where the middle Eocene deposits are lacking. Such an assumption is based on the presence of fragments of middle Eocene nummulitic limestones in the upper Eocene "Asuretia Conglomerate," on the one hand, and on drilling data in Azerbaijan (boreholes near Damitrapa and Sadzhag), where synchronous deposits are represented mainly by marl, on the other.

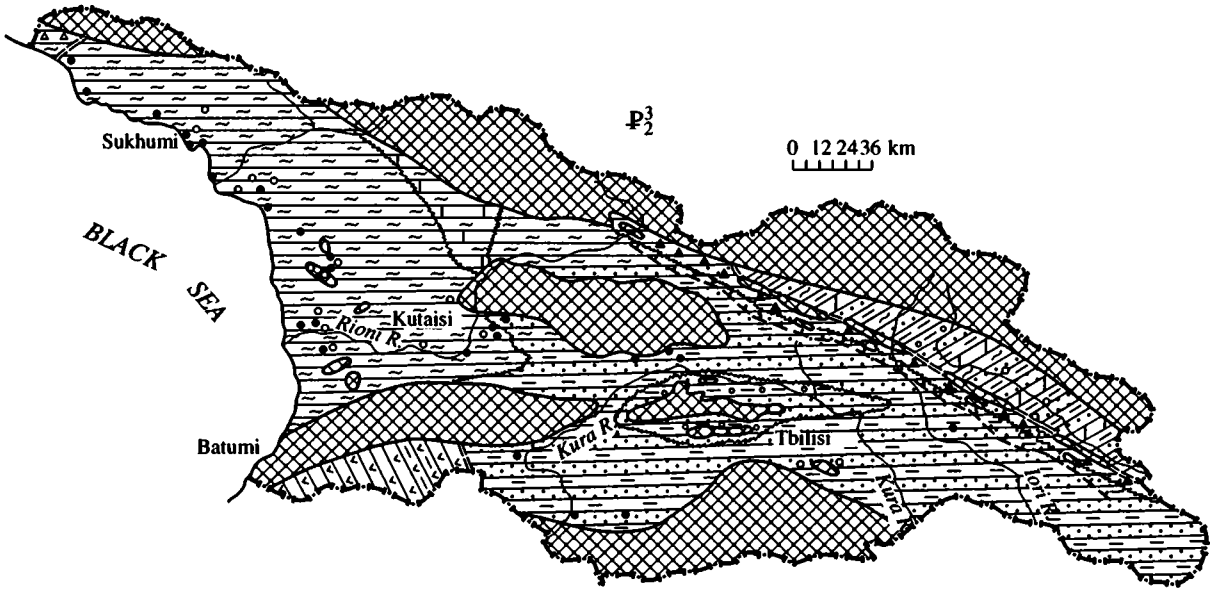

Fig. 6. Paleogeographic and lithofacies scheme of Georgia for the late Eocene (compiled by Maisadze; symbols as in Fig. 3).

In the late Eocene (Fig. 6), the Trialetian (Paleopyrenean) and Neopyrenean (Pyrenean proper) phases of folding resulted in considerable changes in the paleogeographic and facies patterns in the region and throughout the Alpine fold system.

Geologic history of flysch and epicontinental basins, where tectonic settings and facies conditions changed throughout the late Eocene, is considered below, separately, for the first and second halves of the late Eocene.

In the first half of the late Eocene, the accumulation of flyschoid deposits was still in progress in the Western flysch basin. Barren deposits (mudstone, sandstone, marl) resting on the middle Eocene rocks, dated by fauna (the right slope of the Pshitsa River gorge), can conventionally be assigned to this age.

Continuous sedimentation persisted in the Eastern flysch basin. The sedimentary sequence begins here with aleurolitic flysch (the lower part of the Ildokani Formation) consisting of dark deposits (mudstone with rare interlayers of marl) and turbidites (siltstone with sandstone interbeds). Their thickness averages 300–400 m.

The late Eocene transgression resulted in partial increase of sedimentation areas of the epicontinental basin, which were situated south of the Eastern flysch basin and separated from the latter by the Rachinsk–Vandam cordillera zone.

The Kuma Formation accumulated in the remaining part of the epicontinental basin, as elsewhere in the Caucasus and Crimea, at the beginning of the late Eocene. The formation consists of peculiar facies of *Lyrolepis* Marl (the Egris Formation), whose age is still debatable. On the basis of regional tectonic considerations, I refer the Egris Formation to the lower part of the upper Eocene.[1]

The deposition of *Lyrolepis* Marl shows that conditions of sedimentation between the middle and late Eocene changed substantially, and sea basins appeared to be contaminated with hydrogen sulfide. A wide geographic range of the Kuma Formation indicates the regional scale of facies changes. It may be assumed from these facts that a temporal and partial closure (or deteriorated connections with the open sea) of the Crimea–Caucasus basin at the beginning of the late Eocene, which caused its contamination with hydrogen sulfide, was a consequence of the Trialetian phase of folding. A similar but much stronger contamination of the basin occurred in the Oligocene.

Soon after the deposition of *Lyrolepis* Marl, when connections between the basin and open sea had been restored completely to give rise to normal sedimentation conditions, the Kldiani Formation (in the eastern margin of the Adler depression) and the Argveti Formation (in the main part of the basin) began to form.

[1] *Editorial Board remarks:* Author believes that the Kuma Formation and its age analogs in Transcaucasia, the *Lyrolepis* beds (the Egrisi Formation), are the late Eocene in age. However, the middle Eocene (Bartonian) age of the clay–marl sequence that accumulated under conditions of the hydrogen sulfide contamination had been reliably proved for the Crimea–Caucasus region. The regional tectonic considerations cannot be taken into account in this case, since the time scale has its own basis. The Kuma Formation was referred to the upper Eocene in the prewar time and shortly after the war, when this unit included the Bartonian, to which the Kuma Formation belongs. When the Paleogene stages were revised by the International Subcommission on the Paleogene, two of them, the Lutetian and Bartonian, were incorporated into the middle Eocene.

The second half of the late Eocene was marked by the onset of the Neopyrenean orogeny, which peaked in the period of the formation of the olistostromes and "horizon with inclusions" of the Matsesta Formation.

During this time interval, the closure of the Western, and, somewhat later (at the end of the late Eocene), of the Eastern, flysch basins was in progress on the southern slope of the Greater Caucasus.

Beginning from the second half of the late Eocene, the Western flysch basin (the Psou and Bzyb river heads) was transformed into the erosion area that, along with the Akhtsu–Katsirkha block, represented a part of the West Island of the Greater Caucasus.

The lithologic and mineralogic studies of the Paleogene deposits of the Abkhazia zone (Maisadze, 1987) showed that the island was mainly composed of Mesozoic carbonate and volcanogenic deposits; its Paleozoic crystalline and metamorphic rocks were exposed only in the Main Caucasian Range. Staurolite occurring in the upper part of the Paleogene deposits (the Matsesta and Khosta formations) indicates the erosion of the Paleozoic Lashtrak Formation (the Labin Group), which rich in this mineral.

A considerable mass of terrigenous material has been transported into the Eastern flysch basin since the second half of the late Eocene to be accumulated in the regressive upper part of the Ildokani Formation (800–1100 m thick). The formation is characterized by a rhythmic alternation of turbidites (65%) and background deposits (35%).

The distal facies zoning was characteristic of the flysch basin.

Turbidites dominated over pelagic (background) deposits in the souther part of the basin. Aleurolitic–psammitic flysch with an admixture of carbonate accumulated here. Coarse flysch corresponding to conglomerate–breccia horizons (the Aleura and Medzhuda rivers) was also deposited sometimes. Turbidites accumulated in a lesser amount (up to 40%) in the northern part of the basin, where the Ildokani Formation consists of flyschoid deposits.

Judging from these distribution patterns, it may be inferred that the terrigenous influx to the flysch basin was from the Rachinsk–Vandam cordillera zone, situated in the south. Small erosion areas existed within the basin as well; they were responsible for the accumulation of coarse flysch beds during the periods of intense tectonic processes responsible for the formation of olistostromes.

In the epicontinental basin, the Neopyrenean orogeny was most pronounced in its northwestern part (the eastern margin of the Adler depression) and also in the Rachinsk–Vandam Cordillera zone. In the former area, an intense subsidence of the basin floor and formation of the foredeep, where the Matsesta Formation accumulated, were in progress since the second half of the late Eocene. Lithologic features and large thickness (350–400 m) of the formation, as compared to the underlying rocks (the Kildiani and Egrisi formations, 40–70 m), indicate convincingly its regressive nature. Abundant fauna and the well-defined age of separate horizons of the Matsesta Formation, which is synorogenic, elucidate the nature and time span of the Neopyrenean folding (Maisadze, 1984).

The Gagra submarine uplifts located south of the Adler depression controlled at that time the distribution of terrigenous material between the foredeep formed recently and the epicontinental sea with terrigenous and carbonate sedimentation (the Gudauta depression).

The deposition of foraminiferal marls (the Argveti Formation) continued in the main part of the epicontinental basin. Sandy glauconite limestones with interlayers of glauconite sandstones (Agva Horizon with glauconite content up to 70%) accumulated only in the western part of the Rachinsk–Lechkhumi syncline, which represented a shallow basin in the latest Eocene.

The Neopyrenean orogeny was most intense in the Rachinsk–Vandam cordillera zone, where olistostromes discontinuous in the lateral direction formed in the second half of the late Eocene, when the orogenic activity was at the peak. I support the opinion that the olistostrome formation was related to the onset of thrust tectonics in the southern slope of the Greater Caucasus (Leonov, 1975, Gamkrelidze, 1977). However, contrary to these and other authors, I do not consider some of these olistostromes (east of Bolshaya Liakhvi River) as the "wild flysch" and believe that the upper Eocene olistostromes of the southern slope of the Greater Caucasus were formed under paleogeographic and structural environments similar to those in the northern peripheral part of the Gagra–Dzhava zone of epicontinental sedimentation (Maisadze, 1994).

The nappe formation along the frontal thrust contact of flysch deposits resulted in destruction of the Rachinsk–Vandam cordillera zone and in the transportation of disintegrated material in a southward ditection toward the epicontinental sea. These olistostromes, in my opinion, are synchronous to the "horizon with inclusions" of the Matsesta Formation. They mark the culmination of the Neopyrenean folding, which lasted 2–3 m.y. in the second half of the late Eocene (Maisadze, 1984).

On the basis of the abundance and formation mechanism, the upper Eocene olistostromes of the Alpine fold system (Leonov, 1981) should be referred to "event-related deposits."

The Pyrenean orogeny degraded after the accumulation of the olistostromes, to give way to normal sedimentation environments, when the upper portion of the Matsesta Formation (clayey–sandy horizon) and sediments with normal inclusions above accumulated the olistostromes. The latter are usually overlapped by flysch deposits, thrust over them, and are preserved only in individual sections (Aleura and other rivers).

In the Adzharia–Trialeti rift basin, rifting ceased to give way to compression and differentiated movements

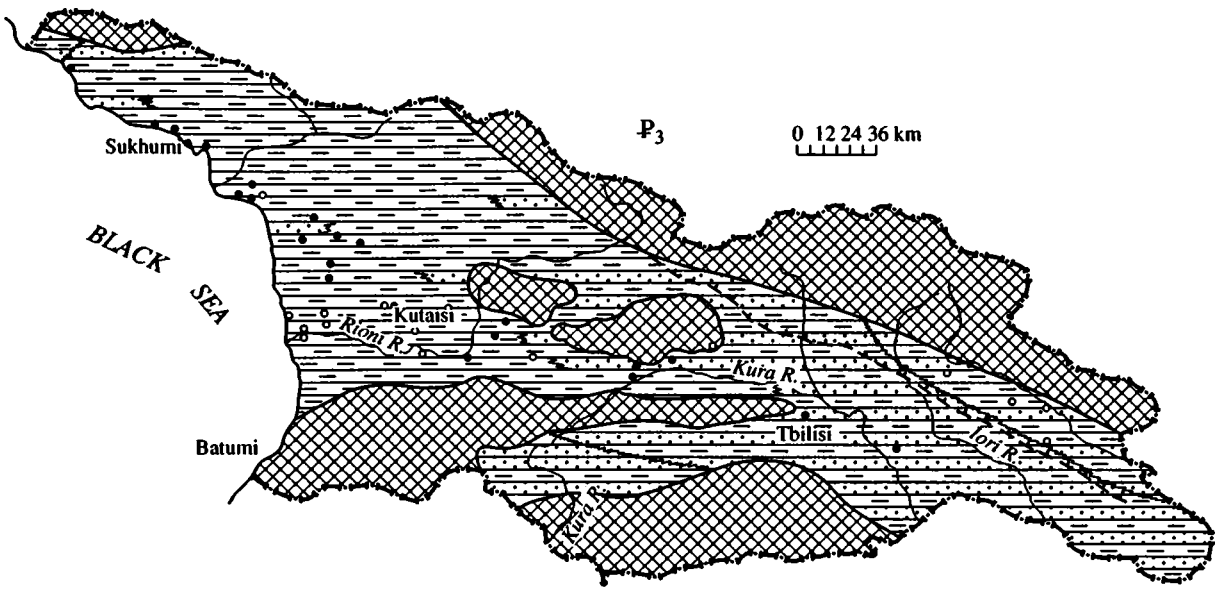

Fig. 7. Paleogeographic and lithofacies scheme of Geogia for the Oligocene (compiled by Maisadze; symbols as in Fig. 3).

in the late Eocene (Gamkrelidze, 1976). This resulted in a sudden attenuation of volcanism in the basin.

The differentiated tectonic movement of the late Eocene transformed the central part of the western segment of the basin into a land that separated the rift basin into two parts (Fig. 6). The sedimentation was still in progress in the Adzharia and Akhaltsikhi depressions on the south, in the Guriya depression, and along the northern periphery of the Adzharia–Trialeti zone in the north.

Volcanogenic–sedimentary deposits of the Adigeni Formation (west of the Adigeni) accumulated within southern Adzharia and the Akhaltsikhi depression at the beginning of the late Eocene. Later on, intensified volcanism resulted in deposition of the Nadadigeni Formation (2000 m thick). Carbonate–terrigenous deposits (110–550 m) accumulated in the eastern part of the Akhaltsikhi depression during the late Eocene.

Lyrolepis and foraminiferal marls accumulated in the main part of the Guriya depression at that time.

The eastern part of the Adzharia–Trialeti basin retained its previous boundaries in the late Eocene; however, the Trialeti cordillera was exposed again in the inner part of the basin. *Lyrolepis* Marl, along with sandstone and clay beds enclosing conglomerate interlayers, accumulated north of the cordillera, whereas south of it bituminous clays with interlayers of graywacke sandstones were deposited at that time.

In the *Oligocene* (Fig. 7), paleogeographic changes resulted in peculiar depositional conditions and caused accumulation of the Maikop Group.

As indicated above, the Eastern flysch basin has been closed since the beginning of the Oligocene, and its structures joined the Eastern Island of the Greater Caucasus. The rise of the flysch trough probably was responsible for ingression, and sea water covered the Rachinsk–Vandam cordillera zone, where the Oligocene–lower Miocene Kinta Formation accumulated. Its deposits 300–400 m thick are represented by the alternation of loose sandstones and the Maikop-type clay.

Peculiar environmental conditions arose in the epicontinental basin, with the exception of its northwestern part (east of the Adler depression edge). A semiclosed basin (Fig. 7) was formed in the central part of the Tkibuli–Sachkheri land in response to transgression. Sandstone beds that accumulated here underlie the ore horizon of the Chiatura manganese deposit represented by siliceous zeolite tuffs and tuffites. The ore horizon is overlain by clayey to spongolite sandstones, which are gradually replaced eastwards by clay (Dzotsenidze, 1965).

The same replacement of siliceous deposits occurred west of the Tkibuli–Sachkheri land in the eastern Mingrelia.

Omitting problems concerning the origin of the Chiatura manganese deposit, which are very debatable, I should point out that the new paleogeographic landscape and partially closed basin were favorable for concentration of manganese in a small area, and this was one of decisive factors responsible for formation of the Chiatura deposit.

The earliest Oligocene was marked in the most part of the epiconinental basin by further downwarping of the depressed parts and by formation of starved troughs, where the continuous sedimentation was in progress (the Gudauta and Ochamchire depressions, the Kura–Iori interfluve, and other areas). The Khadum Formation of carbonate clay (the carbonate content decreases up-section) with sandstone and marl interlay-

ers accumulated at that time under conditions of normal salinity. In terms of lithology, the formation is persistent and represent the reference horizon. Its thickness exceeds 70–80 m.

A peculiar fauna of phosphorescent fish (Danilchenko, 1960) inhabited separate areas of the basin (Gumista River). These fish live in modern basin below the 1000-m depth.

It may be inferred on the basis of the peculiar features of the Maikop sea that conditions favorable for this ichthyofauna could exist at a shallower depth (500–600 m). Judging from the lithofacies nature of the Oligocene deposits and from the geologic considerations, the considerable depth could hardly exist in the area of Gumista River in the earliest Oligocene. The appearance of phosphorescent fish in that area was probably a result of an inflow of a considerable amount of pelitic substance, which caused the lower level of the euphotic zone to rise and fishes to appear at a smaller depth.

The contamination of sea water with hydrogen sulfide and the accumulation of typical Maikop facies, which are represented by bithuminous carbonate-free gypsum-bearing clay beds with sandstone interlayers, took place at the end of the early Oligocene. The deposits contain abundant fragments of fish scales. The maximum thickness of the Maikop Group measures 3000 m (Laliev, 1964).

Facies of the Maikop Group monotonous in the lithological composition indicate their accumulation in a poorly aerated water below the normal basis of wave activity and under stagnant conditions, when the partial or complete contamination with hydrogen sulfide took place. Judging from the biota, geochemistry, and lithology of the Maikop Group, it may be concluded that it has been deposited under conditions similar to the "Black Sea model," which is thought to be a typical example of a closed basin (Hallam, 1983).

It may be also inferred that the accumulation of abundant organic substance in sediments (the result of decomposition of dead organisms on the bottom) under reduction conditions was responsible for the origin of syngenetic pyrite. A slow decay of organic substance in water, contaminated with hydrogen sulfide, determined dark colors of the Maikop Group rocks and of the upper Eocene *Lyrolepis* Marl.

The subsidence, as well as the influx, of sandy material still persisted in the foredeep at the eastern margin of the Adler depression in the Oligocene. This resulted in the accumulation of the Khosta Formation (thick-bedded sandstone 400–450 m thick) and then the Sochi Formation (sandy clay with sandstone interlayers 550 m thick).

Within the Adzharia–Trialeti rift basin, the Oligocene sedimentation was still in progress only in the eastern part, whereas marine conditions were retained only in the Akhaltsikhi (east of the Adigeni meridian) and Guriiskaya depressions (Fig. 7).

Normal sedimentary rocks (coarse-grained sandstone and clay) accumulated at that time in the Akhaltsikhi depression under conditions of a semi-closed basin (Zirakadze, 1982).

The problem of terrigenous material provenance for the Oligocene–early Miocene Maikop basin of the Crimea–Caucasus region is the topical one. The problem requires further investigation and a resolution of alternative viewpoints (Mirchink and Shurygin, 1974; Kopp and Shcherba, 1985; Maisadze, 1987; Stolyarov and Ivleva, 1989, and others). It is hardly probable, in my opinion, that any local provenance in the considered areas could provide the sufficient amount of terrigenous material.

In conclusion, assessing roughly the climatic and hydrologic conditions in the basin and adjacent dry land of the Paleogene time, it may be said that the former represented a sea of the humid zone. Judging from its fauna, marine environs, which existed here in the Paleogene and Eocene, was characterized by normal salinity and corresponded to normal marine or oceanic conditions.

CONCLUSION

The compiled paleogeographic and lithofacies schemes of Georgia for separate epochs of the Paleogene show that characteristic features of geologic history of the region was as follows:

(1) the studied area represented a part of the northern active continental margin of the Tethys, where a system of island arcs, marginal, and interarc basins was developed in the course of a complicated geologic history;

(2) three types of rock associations accumulated in the Paleogene under different geodynamic conditions are represented by flysch, epicontinental–marine, and volcanogenic–sedimentary formations;

(3) the Rachinsk–Vandam cordillera zone, which extends along the southern edge of the Eastern flysch basin of the Greater Caucasus and is completely overlapped nowadays by flysch deposits thrust over it, was the source area for turbidity flows supplying the flysch basin with terrigenous material;

(4) the correlation of olistostromes of the southern slope of the Greater Caucasus with analogous rocks of adjacent areas defines their age as corresponding to latest Eocene; this indicates that olistostromes are synorogenic and synchronous to the "horizon with inclusions" of the Matsesta Formation;

(5) the formation of olistostromes was probably related to nappe thrusting in the southern slope of the Greater Caucasus; the widespread occurrences of these formations within the Alpine fold system allows the upper Eocene olistostromes to be referred to the "event-related deposits";

(6) facies analysis of the upper Eocene deposits in the eastern margin of the Adler depression and tectonic

history of the depression reveal the onset time of Neopyrenean phase of folding that spanned, as established, the second half of the late Eocene lasting 2–3 m.y. in the time interval of accumulation of the Matsesta Formation;

(7) the Pyrenean epochs of the tectogenesis (Paleo- and Neopyrenean phases) were manifested with different intensity in the studied region, as in the whole Caucasus; the Paleopyrenean (Trialetian) phase was most pronounced in the Adzharia–Trialeti zone, whereas the Neopyrenean orogeny was effective in the southern slope of the Greater Caucasus; and

(8) the degree of compression in the Eastern flysch basin of the Greater Caucasus makes up 45–50% for the Paleogene time, while that in subplatform basin of the Ksani–Arkaly parautochthon is equal to 25–30%; the total amplitude of nappe thrusting over the Paleogene section of the southern slope of the Greater Caucasus seems to be 9–12 km.

Reviewer M.A. Akhmet'ev

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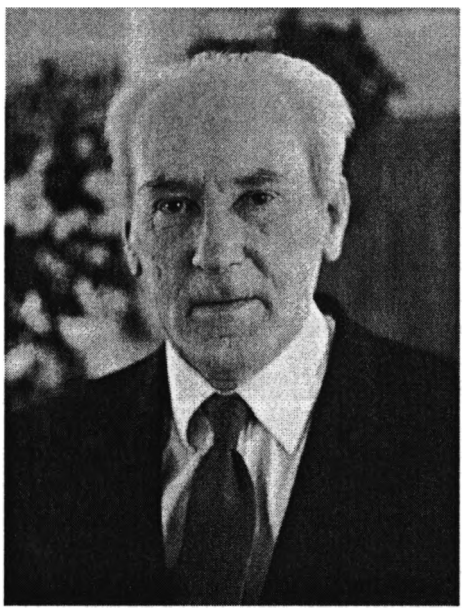
Obituary: Valerii Ivanovich Yarkin (1925–1996)



Valerii Ivanovich Yarkin died on November 18, 1996, at the age of 71. He was a top-level specialist in stratigraphy and paleontology of the Paleogene, one of the authors of the first Russian *Stratigraphic Code* and one of the oldest members in the scientific staff of VSEGEI (All-Russia Institute of Geology, St. Petersburg).

V.I. Yarkin was born on September 15, 1925, to a family of school teachers in the town of Tutaev (formerly Romanov-Borisoglebsk), Yaroslavl Region. At the outbreak of the Second World War, he had just finished school, and he worked almost a year as a turner at the Rybinsk Aircraft Factory. In the spring of 1942, he entered the Aircraft Technical Secondary School in Rybinsk. In later years, he often spoke of his teachers with gratitude, most of whom were high-school professors evacuated from Leningrad. In 1945, he was admitted to the first year course at the Moscow Aircraft Institute, where he studied for only one year and then entered the Geological Faculty of Leningrad University. The change of schools was not spontaneous: even when he was a boy, his chief interests were concentrated on natural sciences.

On graduation from the university in 1951, Yarkin became a postgraduate student at the Historical Geology Department and under the guidance of Professor I.A. Korobkov defended his Candidate's Thesis, *The Lower Paleogene Stratigraphy and Molluscan Fauna of the Turgai Depression and Northern Aral Region.* The material for the thesis was collected during the Turgai Expedition of VSEGEI, when he worked parallel with his university studies. The success of the young researcher was so impressive that Yarkin was soon made Senior Scientist at VSEGEI, and all his later life he was associated with this institute. His research developed along three main courses: the Paleogene stratigraphy and fauna of the vast territory of our country, the general problems of stratigraphic classification and terminology, and geological mapping.

Yarkin successfully worked in the Caspian depression, Volga–Ural area, Ukraine, Caucasus, western Siberia, western Kazakhstan, and Central Asia. In 1961, he published the results of his extensive studies on Paleogene mollusks in the lower reaches of the Volga River and Obshchii Syrt. These materials were used to elucidate, for the first time, the detailed zonation and correlation of Paleogene sections in the entire Volga–Ural region. Under his guidance, a large group of biostratigraphers revised the time intervals and boundaries of the upper and middle Eocene in the European part of the Soviet Union, western Siberia, and Central Asia in publications issued in 1966 and 1973. Another monograph of 1977 described methods of correlation using different faunal and floral groups to obtain a complex characterization of the stratotype and parastratotype Paleogene sections in the southern part of the Soviet Union. An outstanding expert in the Paleogene stratigraphy, Yarkin was the author of the corresponding parts in fundamental monographs *Geological Structure of the USSR* (1958, 1968), *Geological Structure and Minerageny of the USSR,* and *Stratigraphy of the USSR* (1975). He also described many stratigraphic units in the *Stratigraphic Dictionary of the USSR* (1956, 1982). In 1978, Academician V.V. Menner passed on to Yarkin his leadership in the Commission on the Paleogene System of the Interdepartmental Stratigraphic Committee (ISC). In the 1980s, the commission convened large meetings on stratigraphy and organized field sessions (in the Carpathians in 1982, and in other localities) with excursions to the Paleogene reference sections. In 1980–1983, Yarkin compiled the first unified stratigraphic scheme of the Paleogene in the Caucasus and in 1994 completed a comprehensive study of interregional Paleogene correlation in the territory of the Russian platform, the Urals, and western Siberia.

When the Commission on Stratigraphic Classification, Terminology, and Nomenclature of the ISC was set up in 1965, Yarkin became one of the most active members of its bureau. He had two main tasks: the preparation of the first national *Stratigraphic Code* and the chairmanship of the Subcommission on Correlation Stratigraphic Schemes. Yarkin prepared in 1967 the first variant of the code that was extensively used later during the compilation of the final variant. He also served as one of the main compilers and as a member of the editorial boards for both issues of the *Stratigraphic Code* (1977, 1992). His painstaking work on approbation and general appraisal of all Phanerozoic stratigrahic sections that were adopted at the regional meetings is of particular importance. He studied dozens of sections, and his advice to the authors and comments on methods was inestimable.

Yarkin was a member of the ISC bureau and its Permanent Commissions on Central Asia and the Caucasus, on the Neogene system, and on reference sections. He actively participated in regional meetings that took place in different localities ranging from the Baltic region to the Far East; this arrangement of meetings provided a strict compliance with the requirements of the *Stratigraphic Code* during discussions of correlation schemes and preparation of resolutions.

The third aspect of his activity was geological cartography. In the 1970s, VSEGEI and local geological administrations and organizations prepared the Geological Map of the Caucasus, scale 1 : 500 000, under the editorship of Academician D.V. Nalivkin. As the Deputy Editor-in-Chief, Yarkin carried out enormously complicated coordinating work to reach mutual agreement between geologists, some holding opposite views and belonging to different schools of thought. Only his diplomatic talent in handling such discussions and his tactful approach could have brought about the successful completion of the map and its publication in 1976. Yarkin was coeditor of the *Geological Map of the Azerbaidjan SSR* (scale 1 : 500 000; 1975) and coauthor of the *Tectonic Map of the USSR* (scale 1 : 2 500 000; 1966).

Yarkin dedicated much of his effort and time to his work in the Scientific Editorial Council and the Superintendent Group at the Ministry of Geology. Geologists from the local geological departments of Russia and other republics of the former Soviet Union recall with appreciation and gratitude the consultations and lectures he gave at the Courses of Professonal Improvement organized by VSEGEI, where he freely shared with them his experience and knowledge.

At the beginning of the 1980s, Yarkin was invited by the Geological Faculty of St. Petersburg University to lecture on the Cenozoic in the courses *Stratigraphic Practical Work* and *Supplementary Chapters to Historical Geology.* He met the new challenge with enthusiasm and great interest. He meticulously prepared his lectures and applied different pedagogical approaches; his contacts with students and the results of his teaching were always important to him. The lack of formality and of a bleak routine made his lectures, scientific excursions, and seminars extraordinary happenings. Using concrete material on Paleogene or Neogene sections, he showed the general approaches to solution of stratigraphical and theoretical problems in geology.

The range of interests of Yarkin went far beyond geology. His hobbies were photography, amateur films, close observation of animal life, riding his motorcycle, sailing on his yacht, drawing in ink, and, of course, writing books.

Valerii Ivanovich Yarkin was a kind and friendly man, inherently methodical and with a philosophical turn of mind. His sense of responsibility for any kind of work was extreme, and his encyclopedic knowledge made him an unfailing authority to a wide range of specialists.

Valerii Ivanovich Yarkin will remain in the memory of all who knew him, worked with him, and acquired knowledge from him as a universally recognized paleontologist and geologist and as a talented and generous personality.

B.S. Sokolov, A.I. Zhamoida, B.A. Borisov, L.A. Panova, V.A. Prozorovskii, K.E. Yakobson

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Stratigraphy and Tectonic Evolution of Southeast Asia and South Pacific (International Conference, Bangkok, Thailand, August 19–24, 1997)

Yu. D. Zakharov

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The International Conference on the Stratigraphy and Tectonic Evolution of Southeast Asia and South Pacific took place in Bangkok (Thailand), August 19–24, with 430 participants representing 23 countries: Australia (6), Great Britain (7), Vietnam (6), Germany (6), Indonesia (1), India (1), Iran (1), Cambodia (2), Canada (1), China (7), Laos (5), Malaysia (3), Mayanmara (4), New Guinea (2), New Zealand (4), Papua (2), Russia (1), Singapore (1), USA (3), and Thailand (the overwhelming majority of participants). The conference, which was held in *Siam-City* Hotel, was organized by the Geological Survey of Thailand, the Mineral Researches Department. P. Attavipat, and the Minister of Industry, opened the session.

Contributions on general problems in the geology, environments and mineral resources were presented by invited scientists: N. Mantajit (the Geological Survey of Thailand), W.S. Fyfe (University of Western Ontario), J.A. Grant-Mackie (the Auckland University, New Zealand), J.B. Waterhouse (*Oemaru* Co., New Zealand), M. Vilaihongs (the Department of Geology and Mines, Laos), K. Faungswasdi (*Central Plaza* Co., Bangkok), J. Hinton (*Taxeco Exploration* Co., Thailand).

Principal papers of the session were devoted to three issues: (1) stratigraphy and paleontology, (2) tectonics, and (3) economic problems in extracting oil, gas, and other natural resources.

Thirty papers were devoted to stratigraphy and paleontology. The paper by H. Fontaine (France) and S. Solyapogse (the Geological Survey of Thailand), which was not presented orally, noted the poorly studied Carboniferous deposits of Thailand. Data on the Visean–Serpukhovian diverse coral assemblage were presented for the first time. J.G.S. Goodall (*Nordwich* Co., Great Britain) with coauthors reported on the Early Carboniferous palynoflora of Thailand. T. Ishibashi and M. Fujikawa from the Kyushu University, N. Nakornsri from the Geological Survey of Thailand, G.R. Shi with coauthors from the Dickin University, Australia, presented ingenious data on the Late Paleozoic biostratigraphy, ammonoids and brachiopods of Thailand. K. Shashida (the Tsukuba University) and N. Nakornsri (the Geological Survey of Thailand) showed the importance of the Early Permian radiolarians from cherts of the Khanu Formation of Thailand for paleogeographic reconstructions. The paper presented by Yu.D. Zakharov with coauthors (the Far East Geological Institute, Russia) was dedicated to the detailed correlation of the Permian deposits in Primor'e (Russia) and Kitakami (Japan), which are based on recent paleontological and geochemical data. The Permian time in the Sea of Japan region (Primor'e and Kitakami prefecture) was marked by five events distinguished on the basis of the high content of heavy carbon isotope in organogenic carbonates: (1) the Artinskian–early Bolorian (4.7‰), (2) the early Midian (up to 4.5‰), (3) the middle Midian (3.2–3.7‰), (4) the late Midian (3.2–4.1‰), and (5) the early Dzhulfian (4.0‰). The events are attributable to the high bioproductivity stages in the Far East basins related to the transgressions and warm humid climate phases. It has been suggested that the boundary between the Midian and Dzhulfinian stages in Kitakami prefecture is located 7 m below the top of the Kanokura Formation, where bryozoans characteristic of the Lyudyanzin Horizon (the Dzhulfian) in the southern Primor'e have been indentified. H.J. Campbell (the Institute of Geologic and Nuclear Sciences, New Zealand) presented original data on bivalves, the largest in the world, which were encountered in the Middle Permian deposits of Thailand. These giant shells studied in detail may be up to 2 m wide and 6 m long. Judging from the paper presented by V. Khuc (the Geological Survey of Vietnam) and J.A. Grant-Mackie (the Auckland University), new Bajocian mollusks have been revealed recently from the lower part of the Langa Formation in southern Vietnam. The paper by T. Jinnan (the China University of Geological Sciences) was dedicated to the cyclostratigraphy of the Griesbachian (Lower Triassic) clayey–carbonate deposits from the Meishan section of southeastern China, which recorded the climatic cycles between the Permian and Triassic. P. Chaodumrong (the Geological Survey of Thailand) and C. Barret (the Tasmanian University) presented data on lithofacies in the Upper Triassic deposits of the Fa Daen Formation of northern Thailand and on paleocurrents. The paper by E. Buffe-

taut and his colleagues (the Burgon University) was focused on Late Jurassic and Early Cretaceous dinosaurs, newly found in northeastern Thailand which suggests their migration from Europe. V. Pisutha-Arnond and his co-authors (the Chulalonkor University) reported data on distribution of sulfur and oxygen isotopes in the Fichit-Nakhon Savan Formation of Thailand composed of the Mesozoic gypsum–anhydrites. Judging form the paper by Lei-Zuo-qi (the Scientific and Technological Research Center, China), which have not been announced, the palynologic succession of Tertiary formations in southern China reflects climatic conditions of that time. According to these data, the strongest cooling event was in the accumulation time of the lower Miocene Zhuhai Formation. The climate became warmer in the middle Miocene time of accumulation of the Hanjiang Formation. The paper by A.M. Yakzan (*Setangor*) was dedicated to the early Eocene climate elucidated by palynologic data. H. Jiamao with coauthors (the Institute of Geology, Beijing) presented materials on the bio- and magnetostratigraphy of loess deposits of the Quaternary Malan, Lishi and Wucheng formations. General problems on the tectonic evolution of southeast Asia were discussed by I. Metcalfe (the University of New England, Australia), S. Hada (the Kobe University, Japan), K. Wakita (the Geological Survey of Japan), and several other researchers. According to data on radiolarians, presented by Metcalfe, the Paleotethys opened in the Middle–Late Devonian and was closed in the Triassic. Among papers related to oil and gas deposits, those of K. Fuangswasdi, P. Chiaprasert (*Central Plaza*, Thailand), R.H. Findlay, and V. Pisutha-Arnond (the Geological Survey of Thailand) were of the greatest interest.

The conference has summed up recent achievements of geological research in vast expanses of southeast Asia and South Pacific and has outlined a way for further work to be performed on problems of stratigraphy, paleontology, tectonics, and mineral resources.

Reviewer M.A. Semikhatov

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GEOLOGY OF ORE DEPOSITS was founded in 1959. Since 1993 the journal has been published simultaneously in both Russian and English. The journal is one of a few special periodicals devoted to metallic minerals and the conditions of their formation and distribution. Since the advent of glasnost, many restrictions on publication of geological data have been lifted, making available more complete and concrete information on important deposits throughout the former USSR, including such unique ore districts as Noril'sk, Udokan, and Muruntau. Also included are articles submitted by foreign contributors on strategic deposits throughout the world.

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